UNION GÉODÉSIQUE ET GÉOPHYSIQUE INTERNATIONALE
INTERNATIONAL UNION OF GEODESY AND GEOPHYSICS

ASSOCIATION INTERNATIONALE
D'HYDROLOGIE SCIENTIFIQUE

INTERNATIONAL ASSOCIATION
OF SCIENTIFIC HYDROLOGY

ASSEMBLÉE GÉNÉRALE DE TORONTO

3-14 SEPT. 1957

GENERAL ASSEMBLY OF TORONTO

TOME IV

NEIGES ET GLACES

VOLUME IV

SNOW AND ICE

PUBLIÉ AVEC L'AIDE FINANCIÈRE DE L'UNESCO

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SECRÉTAIRE : L. J. TISON
RUE DES RONCES, 61, GENTBRUGGE (BELGIQUE)
GENTBRUGGE 1958

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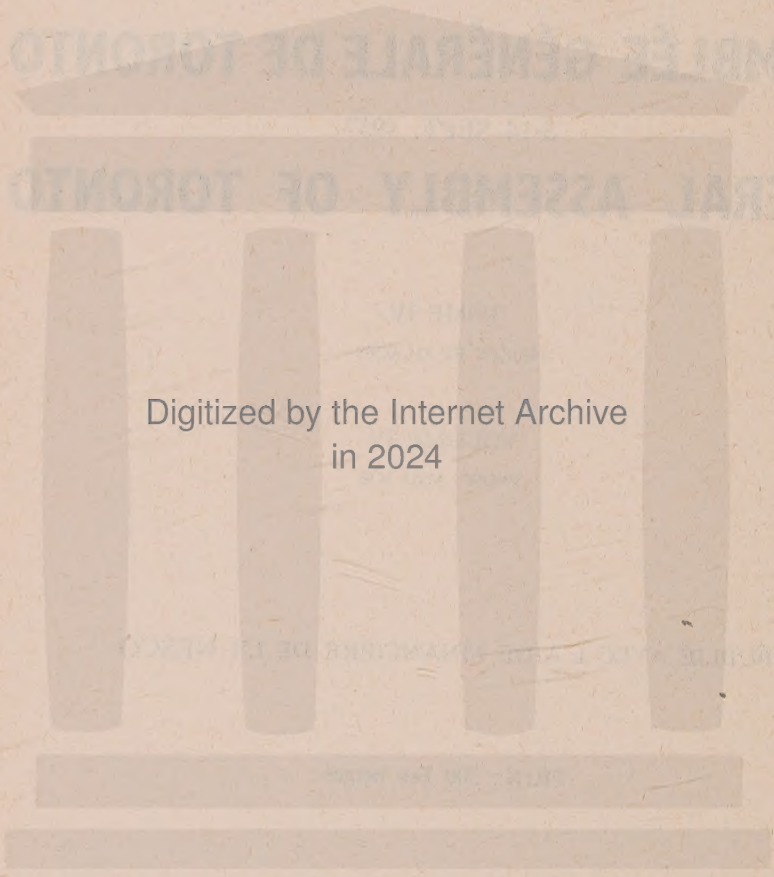

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COMMISSION ON SNOW AND ICE INTERNATIONAL ASSOCIATION OF HYDROLOGY U.G.G.I. ELEVENTH CONGRESS, TORONTO 1957

REPORT OF PROCEEDINGS

FIRST SESSION, 6TH SEPTEMBER, 09,00

Vice-President, Professor R. Finsterwalder in the chair and 45 members present.

1. The chairman prefaced his remarks with an expression of regret at the absence through ill health of the commission's president, Professor R. Haefeli and proposed that greetings and wishes for a speedy recovery be sent to him. This was heartily endorsed by the meeting.

2. Further sad news—the chairman said—was the very recent death of Dr. Harald Sverdrup, one of our best known and most active workers,a nd a world-wide authority on sea ice and related subjects.

3. The chairman apologised for the lack of a presidential address, but said that he had been busy both in Greenland and in the Alps during the period before the congress when it was known that Professor Haefeli was too ill to attend.

4. The secretary (Mr. Baird) read extracts from recent reports from several countries on their glaciological activities in connection with the I.G.Y. Professor Finsterwalder added news on German activities and Dr. H. Bader drew attention to the recent loss of two glaciologists working for the I.G.Y. programme on the Greenland ice cap.

5. A report on progress towards the International Greenland Glaciological Expedition (E.G.I.G.) prepared by Vice-President A. Bauer was read by Dr. de Quervain. This report appears below.

Dr. de Quervain added some information on glaciological work accomplished in central Greenland and Dr. B. Brockamp spoke on the seismic part of the programme. The difficulties of the geoditec profile measurement across the ice cap were mentioned by the the chairman and Dr. T.J. Blachut spoke on recent refinements in radio-altimetry and its possible application to Greenland where the isobaric levels might be very uniform.

6. Professor G. Manley read his paper —

Studies of the frequency of snow fall in Great Britain 1868-1956

Dr. H.H. Lamb in commenting remarked that the correlation between English and Alpine snow fall appeared good, but between England and Iceland — poor. Professor Manley maintained however, that spring snows in England were related to arctic air outflow and could be tied to Icelandic conditions.

7. Dr. A. Bilello read his paper —

A survey of Arctic snow cover properties as related to climatic conditions.

8. Mr. J.G. Potter read his paper —
Mean duration and accumulation of snow cover in Canada.

SECOND SESSION, 6 SEPTEMBER, 14.00

Professor G. Manley in the chair and 50 members present.

1. Professor J. Tuzo Wilson (Vice President of the Union) spoke on the new *Glacial Map of Canada* which is in preparation. Later the manuscript map was displayed and its symbols explained by Mrs. G. Falconer.

2. Dr. L.W. Gold read his paper —
Some observations on the influence of the snow cover on heat flow from the ground in the Ottawa area

Contributors to the following discussion were Thornwaite, de Quervain, Higashi, Manley and Hamilton. The first named stressed the importance of measuring evaporation from the snow surface which Dr. Gold said was now under way, the latter mentioned new experiments at the California Forest Experimental Station on the heat flow and evaporation in forest and open spaces.

3. Dr. A. Higashi read his paper —
Snow survey in Hokkaido

In reply to a question he stated that the power for the radio transmitting the snow depth from radio-active tracer material came from dry cells buried well below ground (and snow) level.

4. Dr. J.A. Bender read his paper —
Air permeability of snow

In the discussion Dr. Bader said that from his experience after a density of .50 to .55 is reached in snow packing, grain crushing tends to take place.

5. Mr. J.A.S. Milne read his paper —
Snow Survey by the Ontario Hydro

THIRD SESSION, 7 SEPTEMBER 09.00

Dr. H. Bader (United States) in the chair and 35 members present.

1. Extracts from the following papers were read by Dr. Higashi —
C. Kojima *Recent studies on the mechanism of snow melting in Japan*
M. Shoda *Study of snow-accretion on wires*

2. Dr. de Quervain read the paper by Professor P.L. Mercanton —
Les cinquante ans d'observations a l'Eismeer par le nivomètre.

3. Dr. G.P. Williams read his paper —
The variability of the physical characteristics of snow cover across Canada

4. Dr. de Quervain read his paper —
Pressure and temperature gradients in snow

5. The secretary read extracts from the paper by V.V. Rakhmanov (U.S.S.R.)
Influence of forests on accumulation and thawing of snow in dependence on meteorological conditions

6. Dr. Bader in the time remaining in this session gave some interesting details on recent S.I.P.R.E. work in north-west Greenland. He described the two deep holes drilled at 2260 m. altitude, 200 miles east of Thule, the second of which has reached 435 m. in depth and from which good core recovery has been made thus retrieving

ice estimated at 800 years o d. Lower core samples however, the bubbles in which reach 22 atmospheres' pressure, are much cracked by the pressure release and he believes that this is a limit of useful core recovery depth. Temperatures range from —24.6° at 20 m. depth to —25.5° at 300 m.

The design of the drill bit and the precooling of the compressed air were also described. The layer of Katmai eruption ash (1912) is traceable and in counting annual layers considerable use is being made of the $0^{16}/0^{18}$ isotope ratio.

FOURTH SESSION, 9 SEPTEMBER 09.000

Dr. M. de Quervain in the chair and 35 members present.

1. Professor Finsterwalder read the paper by Dr. O. Schimpp —
The economy of the Hintereisferner in the years 1952/53, 1953/54.
Dr. Bader commented that the velocity of the wave of maximum velocity recorded (20 km/year) seemed to be the same as that of Alaskan glaciers after the Yakutat earthquake and felt it was important that work should be done on this phenomenon.

2. Professor L. Gherardelli read the paper by Professor M. Vanni —
L'activité du comité glaciologique et les variations des glaciers italiens en 1956,

3. Mr. D.C. Pearce read his paper —
Ground temperature studies

4. Professor Finsterwalder read the paper by W. Hofmann —
The advance of the Nisqually glacier between 1952 and 1956

5. Mr. A. Johnson read his paper —
Investigations on Grinnel and Sperry glaciers, Montana

6. Dr. G.A. Avsiuk read his paper —
Glaciological investigations in the Soviet Union
In answer to enquiries Dr. Avsiuk said that drilling was being done in the Antarctic with a 6 inch core drill using compressed air—that 200 metres depth had been achieved at Pionerskaya, and that bedrock had been reached at 90 m. at Mirny.

7. Professor A. Marussi gave an extract from the two papers by Professor A. Desio —
a. *The Kutiah glacier and its advance in 1953*
b. *Glaciological researches on the Italian Karakorum expedition 1953/55*
and showed maps prepared by the expedition of K_2 and neighbouring glaciers.

8. Mr. R. Carter read the paper by E.P. Collier —
Glacier variation and trends in run-off in the Canadian Cordillera

9. The Secretary announced that Professor P.L. Mercanton had once more completed and submitted his Report on the variations of European glaciers 1954/57 and that this would be published in the proceedings.

FIFTH SESSION, 10TH SEPTEMBER, 09.00

Professor R. Finsterwalder in the chair and 27 members present.

1. This session being, in part, the business session of the commission, the proceedings began with the selection of officers for the coming triennium 1957-60. Dr. de Quervain, who had been appointed to the nominating committee of the

association, said that after consultation with various members of the commission he had arrived at the following suggestions: —

President Professor R. Finsterwalder (Germany)
Vice-presidents Professor A Bauer (France)
Dr. U. Nakaya (Japan)
Secretary Mr. P.D. Baird, continuing.

Those nominations were endorsed by the meeting and transmitted to the association.

2. The Secretary made an announcement concerning the French translation of the book by Dr. P.A. Shumskiy, U.S.S.R.
«Principes de la connaissance structurale de la Glace» available from Centre d'Etudes et de documentation paléontologiques, 3 Place Valhubert, Paris V^{e}.

3. A discussion followed on whether the commission should have set questions for papers to be presented at the next congress, or whether the existing practice of accepting any snow and ice topic should be continued. The difficulties of rejecting any paper submitted, particularly from countries which did not possess any powerful national committee, were stressed. The meeting tentatively approved the suggestion that a small editorial committee should be appointed at the beginning of the next congress period to decide if any papers submitted were not worthy of publication by the association. But the feeling of the meeting also was that no set questions should be posed.

4. It was proposed that attempts should be made, subject to Union approval, to organise a symposium on «The physics of glacier movement» to be held in September 1958 at a suitable locality with good access to glaciers, either at Obergurgl (Austria), Chamonix (France) or elsewhere.

5. It was resolved that messages of greeting should be sent to the Honorary Presidents of the Commission, Drs. J.E. Church and P.L. Mercanton.

6. Further information on the I.G.Y. programme was provided to the meeting by Professor Manley and others. The World Data Centres for Glaciology would be:—

1. The American Geographical Society, Broadway & 156 Street, New York City.
2. Instituta Gidrometeorologichiskiy, Moscow.
3. British Glaciological Society c/o Scott Polar Research Institute, Cambridge, England.

Each data centre is to produce and circulate a catalogue of its unduplicated material, e.g. photographs.

7. Dr. de Quervain reported on the International Snow Classification. He said that in 1956 he had met with representatives of W.M.O. to discuss final changes in the latter's «Cloud Atlas». There were still some small differences between our symbols and those of W.M.O., as shown in the mineographed sheet (which was distributed) but there were more serious differences in definitions. A small working group consisting of himself, Dr. Higashi, Mr. Klein and Dr. Diamond were asked to look into these differences.

8. Dr. de Quervain then gave his paper —
Suggestions for an avalanche classification

9. Miss M. Dunbar read her paper —
Curious open water features in the ice at the head of Cambridge Fiord.
The meeting was at a loss to explain this interesting phenomenon of a late appearing circular hole. Dr. Landauer suggested tentatively that it might be from

a local salt deposit—this, or volcanic activity, would be proved only by a visit to the locality.

10. Professor Finsterwalder read his paper —
The scope, state, and development of precise glacier surveys

In reply to a question by Dr. Bader on the amount of work involved in a photogrammetric examination of a medium sized glacier the speaker said that Hofmann had taken a week to do his survey of the Nisqually glacier.

In connection with the difficulty of height contouring from aerial photographs on a uniform white snow surface, Dr. de Quervain mentioned coal dust being used in Switzerland and Dr. Meier drew attention to the Logetonic process (on exhibit at the Congress) which uses variable light scanning to produce very superior and detail-showing prints particularly of snow covered glacial areas.

11. Dr. Higashi gave a short abstract of the paper by Prof.Yoshida
Physical studies in Japan on the mechanical properties of deposited snow.

Sixth Session, 10th September 14.00

Dr. M. Meier in the Chair and 40 members present.

1. Dr. S. Orvig read the paper by J.D. Ives —
Glaciation of the Torhgat mountains, Labrador

2. Dr. J. Glen read the paper by J.F. Nye, W.H. Ward and himself —
Glaciological studies on Austerdalsbreen, 1955-57

In reply to a question by Dr. Bader the speaker hoped that the difference in velocity (summer/winter) of the icefall would be known at the close of the present season's work.

3. Dr. J.A. Jacobs gave extracts from the two papers by himself and Dr. R.R. Docll and by Dr. W.H. Mathews on the work on the
Salmon Glacier in British Columbia.

The meeting was astounded by the number of deep bore holes achieved in this glacier and the intensity of the seismic soundings.

4. The meeting was adjourned at this point owing to the advancing of the time for Lady Eaton's reception.

Seventh Session, 11 September, 14.00

The President-elect, Professor Finsterwalder, in the Chair and 30 members present.

1. Dr. M.M. Miller read his paper —
Phenomena associated with the deformation of a glacier bore hole

Several members contributed to the discussion after this paper, including Dr. Brockamp who asserted that from the Koch-Wegener investigations on the Storstrømmen (temp. —14°C) where motion was of the order of 1700 m./year, sliding on the bed seemed to make up 95% of the motion.

2. Dr. M. Diamond read the paper by himself and R.W. Gerdel —
Radiation measurements on the Greenland Ice Cap

3. Dr. A. Epstein kindly gave some details on the glaciological application of O^{16}/O^{18} analysis. The ratio of these isotopes in snow is dependant on the temperature

of original condensation hence a clear difference shows between winter and summer snowfall, the sensitivity being 250 times the experimental error.

In answer to questions he explained that a 10 mg sample is sufficient, that 30 samples can be analyzed in 24 hours by mass spectrometer, but the process is costly. It was pointed out that on the Greenland ice cap mixing between summer and winter snow was rare—only the latest fall was involved in blizzard drift.

4. Dr. J.K. Landauer read his paper
On the deformations of excavations in the Greenland névé

5. Dr. de Quervain read the paper by President Haefeli
Mesures des pressions et de déformations dans des galeries de glace.

6. Dr. M.F. Meier read his paper —
The mechanics of crevasse formation
In the discussion which followed it was agreed by all present that 40 m. still seemed to be the maximum crevasse depth observed.

7 Dr. B. Brockamp read his paper —
Recapitulation measurements of seismic reflections on the Pasterze glacier.

8. Dr. G.P. Rigsby read his paper —
He also contributed a brief note on the *Karakoram radiation and heat budget* by Dr. N. Untersteiner who is at present with the American Polar drifting station.
This drew the comment from Dr. Diamond that there seems further confirmation on the slight amount of evaporation on snow fields —from slight in low latitude mountains to negligible in Greenland and Antarctica.

9. Professor Finsterwalder read the paper by O. Pillewizer —
New knowledge about the block movement of glaciers
He also contributed a summary on a paper by Fortsch and Vidal on *seismic measurements in the eastern Alps.*

10. After further brief discussion on the proposed symposium on the «Physics of glacier movement» the meeting closed with a vote of thanks to the Chairman, President-elect Professor Finsterwalder, given by Professor Manley.

P.D. Baird, King's College, Aberdeen.

GLACIAL MAP OF CANADA

G. FALKONER, W.H. MATTHEWS, V.K. PREST AND J.T. WILSON

A map showing the extent of existing glaciers and the chief features of the glacial geology of Canada has been compiled. The map is to be printed in several colours on a scale of 60 miles to one inch, which is the scale used for existing geological and tectonic maps.

The map has been compiled from many sources of information, including reports and maps by the Geological Survey of Canada and provincial departments of mines, published and unpublished reports from universities and companies, maps made by several groups who have examined the air photographs of all the Canadian Shield and some other parts of Canada (over two million square miles in all) and special studies recently made of the Cordilleran Glaciation.

The glacial map displays the patterns of ice flow features, the principle eskers, the disposition of moraines, and the location of some important meltwater channels. The boundaries of the unglaciated regions of western Canada are shown. The ice flow features and eskers form a «complex» of sub-parallel, divergent and convergent ice movement trends that head in linear zones called ice-divides. The boundary is marked between the two complex ice sheets which cover the Cordillera and most of the rest of Canada respectively. Tills of two ages are distinguished in Southern Ontario.

The extent of major glacial lakes and of marine overlap is shown, and some of the better formed raised beaches are indicated.

The examination of air photographs and the compilation have been chiefly supported by the Defence Research Board of Canada and the Geological Survey of Canada. The map is being published by the Geological Association of Canada.

INFLUENCE OF SNOW COVER ON HEAT FLOW FROM THE GROUND

Some Observations made in the Ottawa Area

L. W. GOLD

ABSTRACT

A continuous record was kept at Ottawa, Ontario, of the heat flow from the ground into the snow-cover and temperatures in the ground, snow and air during the two winters 1955-56 and 1956-57. A correlation was found between the difference in the mean air temperature and the temperature 5 centimeters under the ground, the snow depth and the heat flux. Values were obtained for the effective thermal resistance of the snow cover under field conditions.

The earth, acting as a huge heat engine, stores thermal energy in its mantle during the summer months and loses this heat during the winter months. This dynamic process, which is still not completely understood, is determined by a combination of many factors such as geographic location, geomorphic influences and the vagaries of the climate. The result of this process is that the ground undergoes cyclic changes in temperature, the period and amplitude of which depend on the time of year, the weather, the properties of the ground, and the depth below the surface at which the temperature is being measured.

The combination of ground temperature and climate plays a very important role with regard to human activities such as agriculture, forestry and in determining the flora and fauna environment in which we live. Because of the importance of temperature and the relative ease of measuring this property, most investigators, particularly those concerned with the influence of the snow cover, have emphasized this aspect. The Snow, Ice & Permafrost Bibliography ([1]) has listed well over one hundred abstracts of papers dealing with this subject. From temperature measurements, conclusions can be drawn with regard to the apparent thermal conductivity of snow and soil, but as has been pointed out by de Quervain ([2]) and Yosida ([3]) for snow, and Kersten ([4]) for soil, since the thermal conductivity is so sensitive to the physical state of the material, it is very difficult to measure reliably and cannot be expected to stay constant over even short periods of time.

Although it is the temperature level which is most important with regard to human activity and plant and animal life, measurements must be made on the net energy flow out of or into the ground to gain an understanding of how this temperature level is established, particularly in the light of a possible contribution from liquid or vapour movement. To measure the various components of this energy flow at the ground surface through the year is a challenge facing climatologists and others at the present time ([5]).

During the winters of 1955-56 and 1956-57 the heat flow from the ground into the snow cover and temperatures in the ground, in the snow, and in the air were measured at Ottawa (latitude 45°24′ N., longitude 75°43′ W.). These observations are reported here along with correlations that were observed between the heat flow, the air and ground temperatures, and the snow depth.

Apparatus

To measure the heat flow, Gier and Dunkle heat meters, 10 by 10 by 0.1 cm, were coated with wax, encased between two layers of moist clay and buried parallel to the surface about 5 cm under the natural grass cover in an open field. Since the ground in the Ottawa area normally freezes well below this depth, it was thought that energy transfer by liquid water or vapour movement would be negligible and that all the energy transfer would be by normal heat conduction at the 5-cm level, during the winter months. The three heat meters used were located in an area about 1 meter in radius. Adjacent to the heat meter installation, thermocouples were supported at various levels up to one metre above the ground surface on nylon cord stretched between wooden posts placed about one metre apart. The thermocouples were made from No. 28 copper and constantan wire, coated with a white porcelain cement to protect them from the weather and mounted unshielded from radiation. During 1955-56, a temporary thermocouple installation was placed in the ground to measure temperatures down to the 30-cm level. A permanent installation which measured temperatures down to 90 cm was made in 1956 directly below the thermocouples above the ground surface.

The output from the heat meters and the thermocouples were recorded on a Twenty Point Leeds and Northrup recorder located in an instrument tunnel about 200 ft from the observation area. Continuous records were kept during both winters of the snow depth and average snow density.

Discussion of Data

Recorded values of the heat flow were taken from the charts once every two hours and replotted against time. Using a planimeter to find the area under the heat

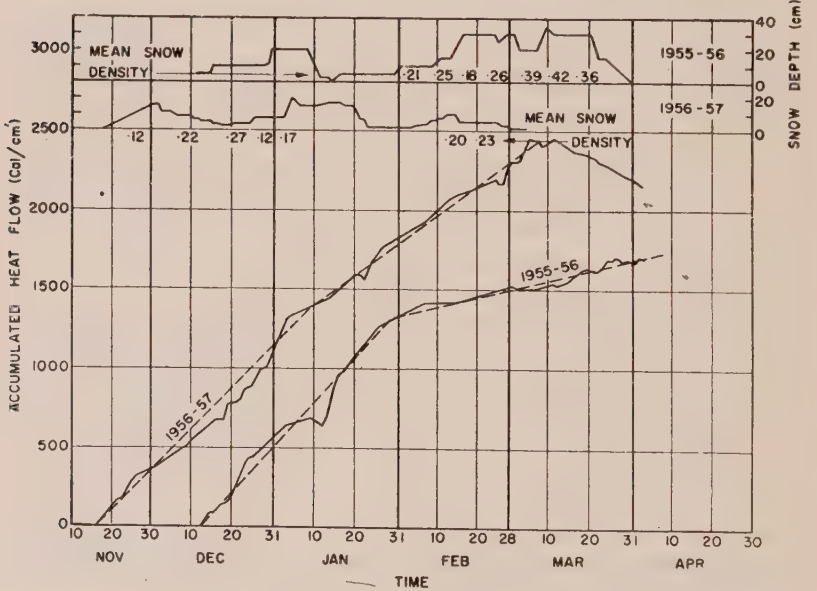

Fig. 1 — Accumulated heat flow, snow depth and density data plotted against time.

flow curves, the total output for each day was found for each heat meter. The mean of the heat meter outputs was taken as the heat flow from the ground to the air or snow cover for each day. A cumulative total of the daily heat flows was then taken, starting from the day on which observations began. This is plotted in Fig. 1. The recorded snow depth and mean snow density are also given.

From 15 to 20 cm above the snow surface to the highest elevation at which the air temperature was measured, the air was generally almost isothermal. The mean value obtained every two hours for this temperature was plotted, and a curve drawn by eye to obtain the daily mean air temperature shown in Fig. 2. A curve was drawn

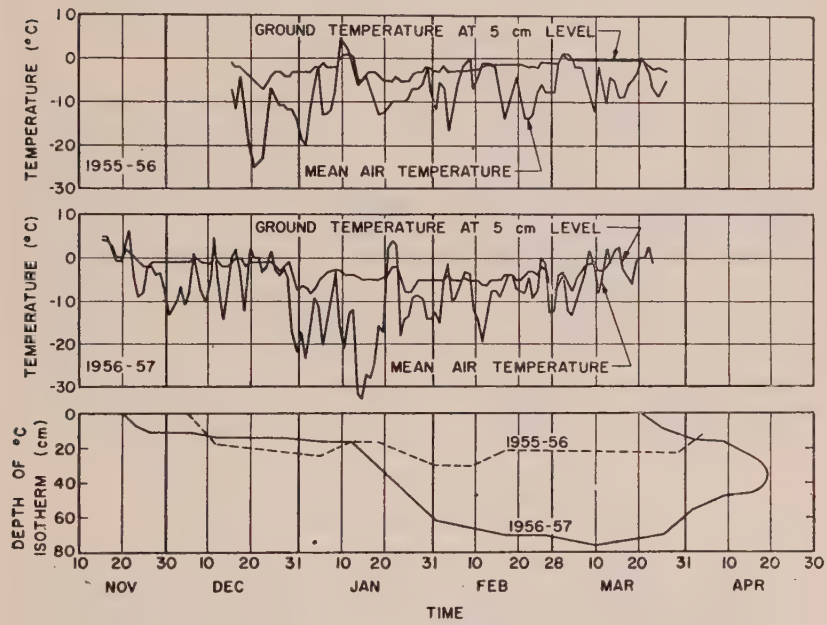

Fig. 2 — Ground and air temperatures plotted against time.

by eye through the two-hourly values for the temperature 5 cm under the ground and this along with the depth in the ground of the 0°C isotherm are also shown in Fig. 2.

From Fig. 1 it is seen that the rate of heat flow, which is given by the slope of the accumulated heat flow curve, is almost constant over extended periods of time. Table 1 gives the average rate of heat flow for various periods as calculated from the slope of the dashed lines.

The weather during the winter in the Ottawa area is quite variable and this is reflected in the snow depth curves. For the 1955-56 season there was an accumulation of snow before 15 January which largely disappeared at this date. After the end of January, the bulk of the winter snow fell and remained until the end of March. In contrast, the initial snow accumulation for 1956-57 almost disappeared after January 20 and very little accumulated after that. These long period variations in snow depth have an effect on the rate of heat flow. The very marked decrease in rate after 15 January 1956 coincides with the major snow accumulation for that year.

The rate of heat flow after 15 January 1957, when there was little snow cover, was not much different from the rate before 15 January. The dips in the heat flow curves at 12 January 1956 and 22 January 1957, coincide with the warm periods responsible for the January thaws. When the cold weather returns the rate of heat flow over a short period is very high, consistent with the shallow cover and the rapid removal of the heat stored in the top layers of the ground during the thaw.

The influence of the mean air temperature on the rate of heat flow is not so obvious although its effect on the ground temperature at the 5-cm level is quite marked particularly when the snow cover is shallow. After 15 January 1956, the temperature at the 5-cm level was not so sensitive to the air temperature and slowly rose to a constant value just below 0°C. During the winter of 1954-55 the depth of snow cover at Ottawa was about 50 cm and was continuous from the end of December until March. Observations taken indicated that in many places there was no penetration of the frost into the ground.

The effect of the snow cover on the depth of frost penetration is also shown in Figs. 1 and 2. It is interesting to note that although the rate of heat flow was almost constant over the winter of 1956-57, the 0°C isotherm remained stationary at about the 15-cm level for over a month and then rapidly penetrated to over 60 cm. This can be explained by a large heat flow from the ground below 15 cm before 10 January which rapidly drops off, or by the formation of ice lenses in the soil at the 15-cm level during the stationary period. Calculations on the thermal conductivity of the soil and observations on the soil moisture content which show a very sharp increase above the 20-cm level indicate the probability of the latter explanation.

Although more heat was extracted from the ground during 1956-57 than 1955-56 over the period 11 December to 11 March, by 3 April the net outflow of heat was almost the same, since the ground began to warm up almost three weeks earlier in 1957 than in 1956. After 20 March 1957, the ground thawed as rapidly from below as from above, a phenomenon which has been reported in the literature ([6]).

Correlation of Heat Flow, Temperatures and Snow Depth

In considering the heat flow from the ground over long periods, the main factors to be considered should be some measure of the mean air temperature and the thermal resistance offered by the snow. The behaviour of a thermal system such as the ground is such that the more quickly the temperature at the surface changes, the more quickly is this change damped out with depth below the surface. Therefore, such short period influences as wind, and radiation, though important in determining the rate of heat transfer from a surface are likely of secondary importance, compared to the much slower changing mean air temperature in determining temperature levels at depth. Indeed, the mean air temperature is dependent on such factors as wind and radiation and to a certain extent a measure of their long term averages.

It was assumed as a hypothesis that the rate of heat flow from the ground was directly proportional to the difference between the mean air temperature and the ground temperature at the 5-cm level, taken from the records every two hours, and inversely proportional to the snow depth. An equation similar to the heat flow equation is thus obtained:

$$Q = \frac{K\Delta T}{X} \qquad (1)$$

where Q is the rate of heat flow cal/cm²hr;

ΔT is the difference between the mean air temperature and the ground temperature at the 5 cm. level in °C;

$X = Xs + 2.5$ in cm,

where Xs is the snow depth and 2.5 is the depth attributed to the soil, assuming that the ground has approximately half the insulating value of snow; and

K is the thermal conductivity of the ground-snow-air combination cal/cm°Chr.

From the temperature plot, the time integral for ΔT over daily periods was obtained graphically. These daily values were than divided by the adjusted average snow depth for the day and the cumulative sum made, i.e.,

$$\int_{\text{day}} \Delta T\,dt \qquad \text{and} \qquad \sum \frac{\int_{\text{day}} \Delta T\,dt}{X}.$$

The accumulated daily heat flows were then plotted against the corresponding accumulated temperature sums as snown in Fig. 3. The dates corresponding to some of the cumulative totals are noted on the curves for each year.

Figure 3 shows that the hypothesis regarding heat flow, temperature difference, and snow depth is reasonably justified over extended periods of time. The greatest variability occurred in the 1956-57 data after the thaw in January when the snow cover was very shallow.

The variability in the 1956-57 data shows that the relationship given by equation (1) is not complete for short-period correlations. The variations in K over short

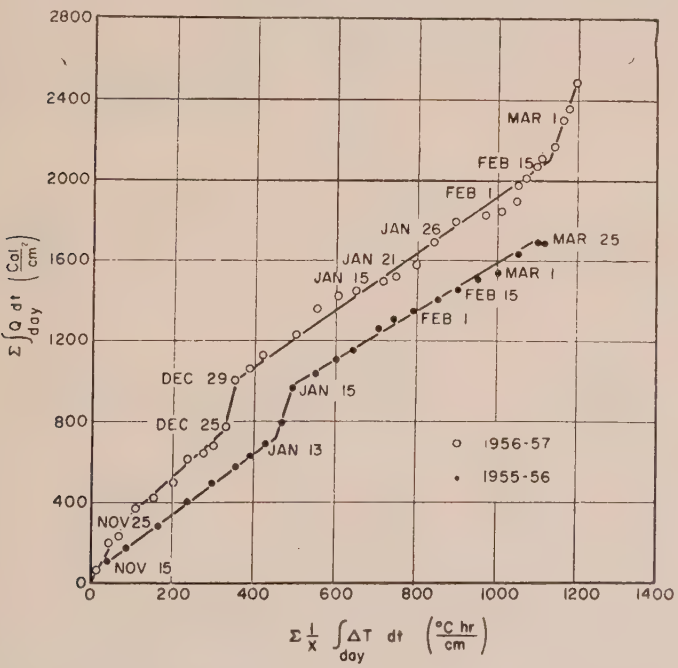

Fig. 3 — Dependence of heat flow from the ground on mean air temperature, ground surface temperature, and snow depth.

TABLE 1

Rate of heat flow from the ground

Year	Period	Average Rate Heat Flow
1955-56	Dec. 10 — Jan. 26	27 cal/cm²day
1955-56	Jan. 26 — April 3	5.7 cal/cm² day
1956-57	Nov. 5 — Jan. 4	25.6 cal/cm² day
1956-57	Jan. 4 — March 7	17.9 cal/cm² day

periods is probably related to the ability of the snow cover to absorb or give up heat energy, particularly through thawing, freezing, and evaporation, during extreme changes in air temperature. The lag that exists between changes in air temperature and changes in the heat flow from the ground would also have an effect. Table 2 gives values of K obtained from Fig. 3 for the time periods indicated.

It is seen in Fig. 3 that during periods when the mean air temperature goes above 0°C, K is greatly increased. One possible reason for this is that, because of the snow cover, the maximum temperature of the ground surface is 0°C and, therefore, the temperature at this boundary is fixed. The heat from the ground does not pass on to the air but is utilized in warming the ground to 0°C and in melting the snow. Thus, the snow cover and the frozen ground act as the sink for heat from both the air and the ground. In this case, the hypothesis set is no longer valid. As can be seen in Fig. 1 and as mentioned earlier, the thaw period affects the rate of heat flow at the time but does not greatly influence the average heat flow over extended periods.

TABLE 2

Calculated values of K

Year	Period	K Cal. cm°C hr
1955-56	Nov. 15 — Jan. 13	1.51
1955-56	Jan. 13 — March 25	1.31
1956-57	Nov. 25 — Dec. 25	1.84
1956-57	Dec. 29 — Feb. 15	1.40
	Mean	1.48

Although the observations for the two winters support the hypothesis, the evidence is still not sufficient to show that the conclusion is generally valid. It is possible that K depends on other factors, particularly climatic, and would, therefore, vary from region to region. From thermal conductivity measurements on snow, K, would be expected to depend on the average snow density, although this dependence is not apparent in the Ottawa observations. Better correlation between heat flow

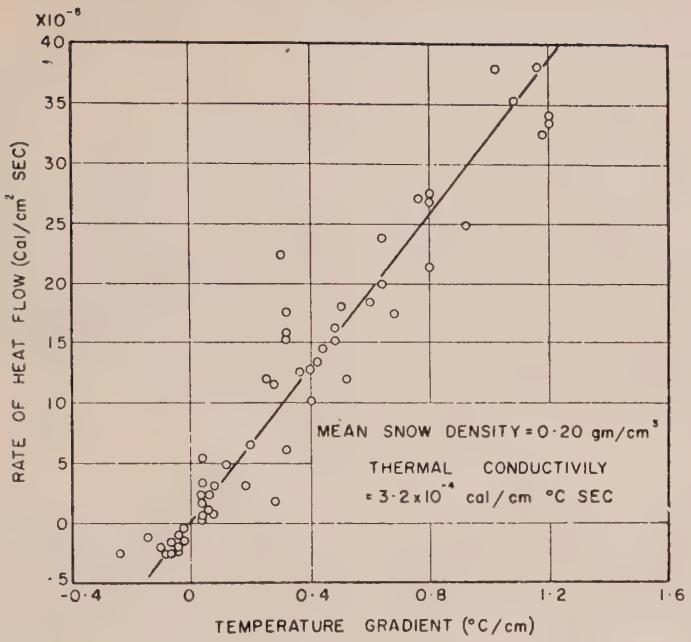

Fig. 4*a* — Rate of heat flow plotted against temperature gradient in lower 10 cm of snow cover (Jan. 31-Feb. 13).

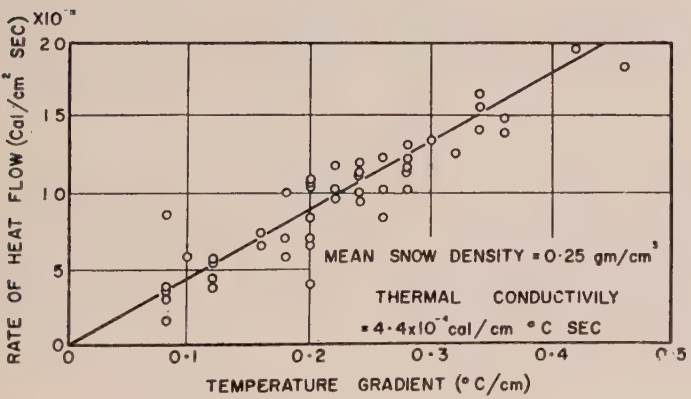

Fig. 4*b* — Rate of heat flow plotted against temperature gradient in lower 10 cm of snow cover (Feb. 16-Feb. 25).

and air temperature may be obtained by using some other measure of the ability of the air to accept or give energy to the ground.

Thermal Conductivity of Snow and Soil

To check the outputs of the heat meters, calculations were made on the thermal conductivity of the snow from the observations of the 1955-56 winter and of the soil from the observations of the 1956-57 winter. Observations on the snow 0 to 10 cm from the ground surface during periods when the heat flow was fairly constant were used for the calculations. The temperature gradient was plotted against the heat flow as shown in Figs. 4*a* and 4*b* and the best straight line fitted by least squares. No attempt was made to correct for the time lag between the heat flow and the corresponding temperature gradient, or for the thermal capacity of the material between the heat meter and level at which the temperature gradient was measured. This probably accounts for much of the scatter of the observations. The measured conductivities are compared with, in Table 3, values given by Sipre ([7]).

TABLE 3
Thermal conductivity values for snow

Period	Mean Snow Density	K Calc. Cal/cm°C sec	K Given (Sipre) Cal/cm°C sec
Jan. 31 — Feb. 13	0.20	3.2×10^{-4}	3.5×10^{-4}
Feb. 16 — Feb. 25	0.25	4.4×10^{-4}	4.5×10^{-4}

The thermal conductivity was calculated for frozen soil to eliminate possible contributions by liquid or vapour movement. It was obtained by calculating the average temperature gradient in the soil 20 to 40 cm below the ground surface over the period 5 January to 24 February and dividing this into the rate of heat flow as obtained from Fig. 1. A value of 0.26×10^{-2} cal/cm °C sec was obtained. The soil was Leda clay with a dry density of approximately 85 lb/cu ft and a moisture content of approximately 30 per cent of the dry density. Kersten obtained a value of 0.40×10^{-2} cal/cm°C sec for soil with similar properties.

From the fair agreement obtained between the observed thermal conductivities and those reported in the literature, it was concluded that the heat meters were giving a reasonable measure of the heat flow from the ground. From observations made on the rate of ice growth on a small pond in a quarry close to the observation area a figure of 23 cal/cm² day was obtained for the rate of cooling. This compares favourably with the rates computed from the heat flow measurements and further supports the accuracy of the heat meters. The heat meters were calibrated in the laboratory using standard hot plate apparatus.

Conclusions

From observations taken in the Ottawa area, it was found that the heat flow from the ground depended directly on the difference between the mean air temperature

measured with an unshielded thermocouple and the temperature at the ground surface and inversely on the depth of the snow cover. A mean value for the conductivity of the ground-snow-air combination was calculated to be 1.48 cal/cm°C hr. It is not considered at this time that the obtained relationship has general validity. The heat flow from the ground in the Ottawa area during the winter is of the order of 20 cal./cm²day.

Acknowledgments

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SNOW SURVEY IN HOKKAIDO, JAPAN

AKIRA HIGASHI
Faculty of Science, Hokkaido University, Sapporo, Japan

Abstracts

Snow survey works carried out in the Hokkaido area by the Hokkaido University group are presented. As a method of snow survey which is suited to natural conditions and urgent needs in Japan, the «quantitative method» was proposed by S. Sugaya. To avoid the disadvantage of time consuming and laborious survey in this method, a practical method was designed for further continuation at the same basin in which the quantitative method was once carried out. For the establishment of the practical method, the snow survey were carried out for four years by the same method in the Chubetsu River Basin. The representativeness of a survey route was proved in various conditions from year to year. This method will be applicable for any basin in which the main factor determing snow deposit distribution is the altitude.

The quantitative method was applied to the Lake Shikaribetsu Basin and the results were used for practical operation of the hydroelectric power plants system which is using the lake as a reservoir. Snow survey works accompanying with more precise studies of the precipitation over the basin are still continuing.

For the efficient use of the water of Lake Shikaribetsu during winter, continuous indication of snow deposit in the basin was required. The radioactive snow gauge was employed for this purpose, and efforts to put it in practical use have been continued for the last few years. Some technical achievements in the construction of the gauge are briefly described.

1. Introduction

The importance of the snow survey wqs recognized in Japan after World War II ended, when she began to utilize her own natural resources more efficiently than before.

The first snow survey in Japan was carried out by S. Sugaya and his colleagues of Hokkaido University at the Chubetsu River Basin in 1948. Since then many snow survey programs have been carried out all over the snow deposit area in Japan under the auspices of electric power companies and a government comittee on natural resources. While these projects have offered very valuable hydrologic data for the operation of hydroelectric power plants and for the construction plans of new reservoirs, they have contributed greatly to the development of survey methods themselves which are best suited to topographic and meteorological conditions in Japan.

This paper is a summary of snow survey works carried out in the last few years in the Hokkaido area by the group at Hokkaido University. The works are composed of three parts, 1) Establishment of practical or simplified quantitative survey methods, 2) Extension of the practical method to the Lake Shikaribetsu Basin and 3) Practical use of the radioactive snow gauge.

2. Establishment of practical snow survey methods in the Chubetsu River Basin

Previous works

On the first snow survey in this basin in 1948, Sugaya [1]) carried out the «quantitative method.» This method differs from the «percentage method,» which is widely

used for the prediction of thawing season runoff in the western United States. By means of this method the total amount of water equivalent to snow deposited over the entire basin is quantitatively estimated from many measurements of snow borings in the area. Thus, the total runoff in the thawing season is directly predictable without any statistical correlation being necessary.

This method is convenient for cases in which the prediction of runoff is urgently needed as in Japan today, because it does not depend on the accumulation of statistical data for a decade of years.

Another reason for employing this method is that we can determine the areal distribution of snow deposits in the basin which are expected to be affected by many factors such as topography, plant cover and meteorological conditions throughout winter etc. The nature of the distribution will provide a rational means for setting up a practical survey method.

We cannot neglect the fact that comparatively smaller catchment areas of Japanese rivers make it possible to carry out this method.

A brief descripiton of the so-called quantitative method will be useful for further understanding of this paper.

The procedure employed by Sugaya is as follows. Snow borings by means of a snow sampler were carried out at 174 points in the Chubetsu River Basin of about 256 km² area. Fifty of them were in the open areas in the forests or on the bare ground of high altitude, and these are used as standard points. The remaining 124 were supplementary points to determine the effects of various factors.

The data from 50 standard points showed very clear linear relationship between the water equivalent of snow deposit and the altitude of the point. This means that a principal factor which governs the charactor of distribution of snow deposit in the basin is altitude, if we disregard the effect of forest cover.

The actual snow deposit in the forest is less than the one in the open areas at the same altitude, and the proportional decrease depends on the degree of forest cover and kinds of trees encountered. Precise comparison of snow deposits under forest cover and in open areas made it possible to calculate a correction coefficient which when multiplied by the open area value would yield the snow deposit value in the forest. Actually in the case of broad-leaved trees, we do not need any correction, e.g. correction coefficieàt is 100%. The correction coefficients for evergreen trees for different densities of cover are shown in Table I.

TABLE I

Correction coefficient for forest cover of ever-green trees.

density of cover	correction coeff. %
open area in the forest	100
thin	90
medium	80
thick	60

Now, if the linear relationship between water equivalent of snow H and altitude h is given as

$$H = a + bh \qquad (1)$$

where a and b are numerical constants, the total amount of water equivalent of snow over a certain area is calculated by the following equation

$$M = R \int H.A.dh$$

$$= R \int (a + bh).A.dh \qquad (2)$$

In this equation, A is the area of a given range of altitude above and below h, and R is the correction coefficient for the forest cover.

In nature, A will not ordinarily appear as an analytical function of h. Therefore equation (2) should be expressed as a summation:

$$M = R \sum_i (a + bh_i).A_i \qquad (3)$$

The basin is divided into several band areas with contour lines of every 200 m and the average altitude of each bands is called h_i, and area of the band between the altitudes $(h_i - 100)$ m and $(h_i + 100)$ m is designated A_i.

In general the correction coefficient R will vary from point to point in the basin, according to the variation in density of forest cover and the ratio of broad-leaved to evergreen trees. Through the inspection on enlarged airphotographs of the basin which were taken at the same time of the survey, the area was divided into three parts in each of which the forest cover condition varied but slightly. This division also mainly depended on altitude, because the ratio of broad-leaved to evergreen trees and density of forest cover are both altitude dependent. These three were designated as high-, middle-, and low altitude forest zones, with some overlapping of altitude as will be shown later.

Besides these three, there are two other zones, namely the zone above the timber line and the cultivated flat zone at the foot of the mountain. The total amount of water equivalent of snow, M_R, in each zone was calculated by means of eq. (3) and the summation for five zones

$$M = M_1 + M_2 + M_3 + M_4 + M_5 \qquad (4)$$

gave the total amount over the entire basin.

By means of the method described above, the total amount of water equivalent of snow in the Chubetsu River Basin on April 1 in 1948 was estimated as 1.94×10^8 m^3.

The value was compared with the total runoff of the Chubetsu River in the snow melting season. Precipitation during this season made this comparison somewhat arbitrary, because of the lack of an appropriate runoff coefficient. However, the total runoff in the snow melting season Q is given as follows.

$$Q = \alpha M + \beta P \qquad (5)$$

where M is total amount of snow, P is total precipitation in the season, α is runoff coefficient of melted snow water, and β is runoff coefficient of precipitation.

Total runoff was measured very accurately at a station which was especially operated for this project, and gave 2.93×10^8 m^3 from April to July. The total precipitation in this season was estimated by multiplying the area by the precipitation in the raingauge which was located at the center of the basin of 1000 m altitude. The proof that the precipitation at this point is quite near the mean value of the precipitation all over the basin was obtained by a study of the precipitation character in the basin for 1948[2]) and the following year[3]). Estimated total precipitation during the season was 1.24×10^8 m^3.

Substituting these numerical values of Q, P and M in eq. (5) and assuming α is 100% to simplify the problem, we can see that β equals 80%. This value of β

seems to be quite reasonable. This means that the total water equivalent of snow coincided with the value which is calculated from the total amount of runoff under the assumption of $\alpha = 100\%$ and $\beta = 80\%$.

Although the quantitative method gives quite an accurate estimate of snow deposition over the entire basin which is useful for direct prediction of runoff, it has the disadvantages of being time consuming and requiring great effort to carry out a complete survey over the entire area. But, when the general topographic character of the snow deposit in the given area is already known from a thorough survey, it will be possible to reduce the number of survey points without greatly reducing the accuracy of the estimate.

Sugaya[1]) has discussed the possibility of a practical method which only requires the determination of the relationship between altitude and water equivalent of snow deposit in the area and a small number of snow boring points. He suggested taking a survey route along the Yukomanbetsu River (a main tributary of the Chubets River) This route traverses the middle of the basin and snow borings at various altitudes up to the highest region of the area are possible.

T. Chiba[4]) carried out a survey along this route in the spring of 1949. Estimation of the total amount of water equivalent of snow over the entire basin in this year did not coincide well with the one which was calculated reversely from the runoff of the Chubetsu River and the precipitation in the thawing season by eq. (5).

In spite of this, the discrepancy can probably be attributed to an insufficient number of boring points above the timber line zone. It seemed worthwhile trying this route again to establish the practical method in this basin.

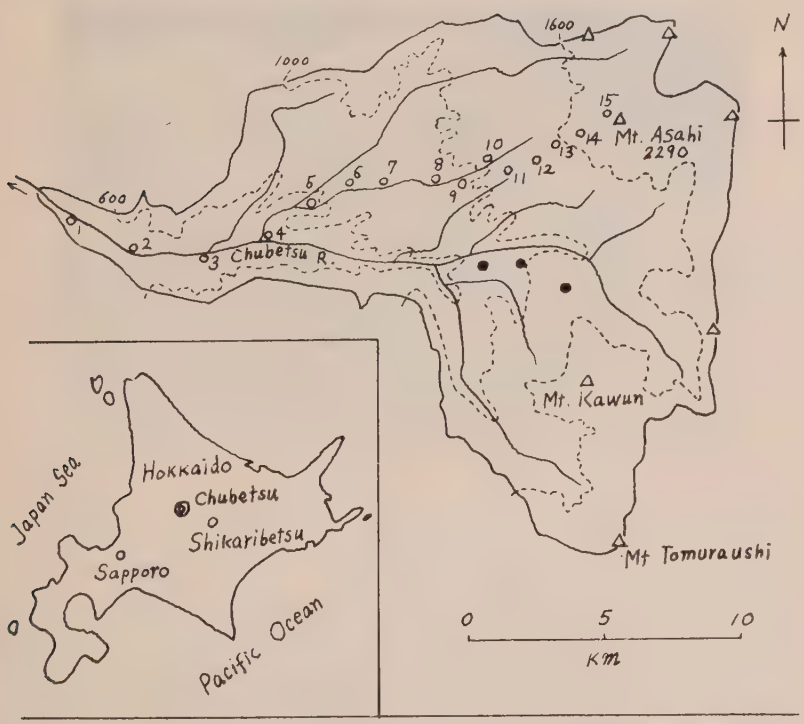

Fig. 1 — General topography of the Chubetsu River Basin.

Fig. 2 — Weighing of bored snow at the site No. 7.

Fig. 3 — Snow boring at the site No. 14.

The snow survey from 1952 to 1954

After two years interruption, the author and his colleagues [5][6]) carried out a three year snow survey to establish the practical method. The 25 km long «Yukomanbetsu survey route» was chosen. This is the same route followed by Chiba in 1949.

General topography of the basin and the survey route are shown in Fig. 1, in which the numbers beside the points correspond to the site number in Table II. Fig. 2 shows the weighing of a snow core in a sampler and the appearance of the

forest at site No. 7 (altitude 760 m). Fig. 3 was photographed at site No. 14 with the top of Mt. Asahi in the background and site No. 15 is shown by an arrow.

Results of the survey are tabulated in Table II.

TABLE II

Results of snow survey in the Chubetsu River Basin in the spring of 1952

zone	site No.	altitude m	depth of snow cm	water equivalent of snow cm
cultivated	1	300	39	15
	2	350	70	29
	3	430	90	36
	4	440	94	40
forest	5	620	149	57
	6	700	156	60
	7	760	192	73
	8	900	225	92
	9	960	199	81
	10	1050	234	93
	11	1200	251	106
	12	1330	297	117
above timber line	13	1510	241	104
	14	1650	210	100
	15	1850	165	57

The data in Table II is plotted on the graph in Fig. 4. As in Sugaya's paper[1]) the relationship between altitude and water equivalent of snow deposit displays different trends above and below the timber line and these two relationships are quite well expressed by linear regression lines. Regression equations obtained by the method of least squares are as follows:

$$H = -4.7 + 0.0952\,h \qquad \text{below timber line} \qquad (6)$$

$$H = 326.7 - 0.1434\,h \qquad \text{above timber line} \qquad (7)$$

In these equations, water equivalent of snow deposit H is expressed in cm, and altitude h is in m.

The standard deviations for these two equations were calculated by means of statistical analysis, and they are shown in broken lines in Fig. 4. Broadening of the width at both ends of each line means that the precision of estimate is low for values remote from the center of altitude range. The standard deviation for eq. (7) is greater than that for eq. (6), because fewer points were obtained in the zone above timber line.

Calculation of total water equivalent of snow deposit based on equations (3) and (4) is tabulated on Table III. Numerical values of area of each band and correction coefficient for each zone were obtained from Sugaya's results.

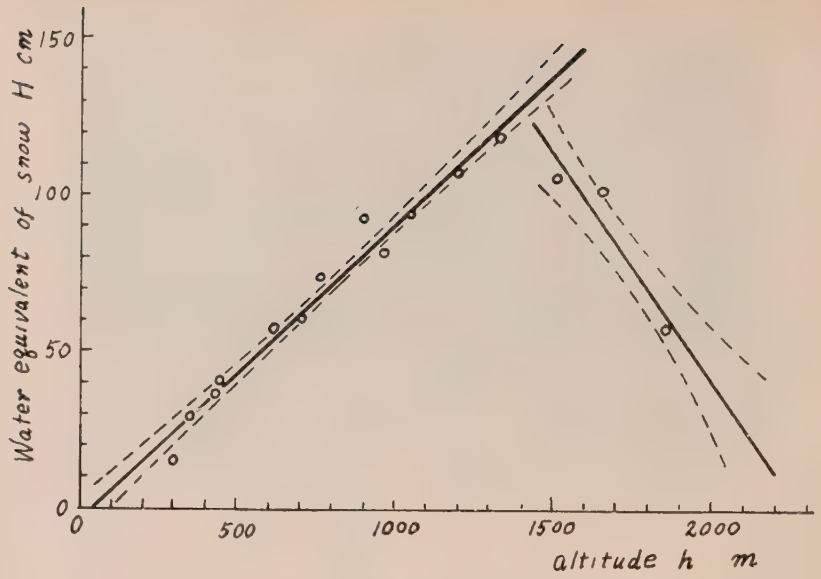

Fig. 4 — Altitude relationship of the water equivalent of the snow deposit in the Chubetsu River Basin in 1952.

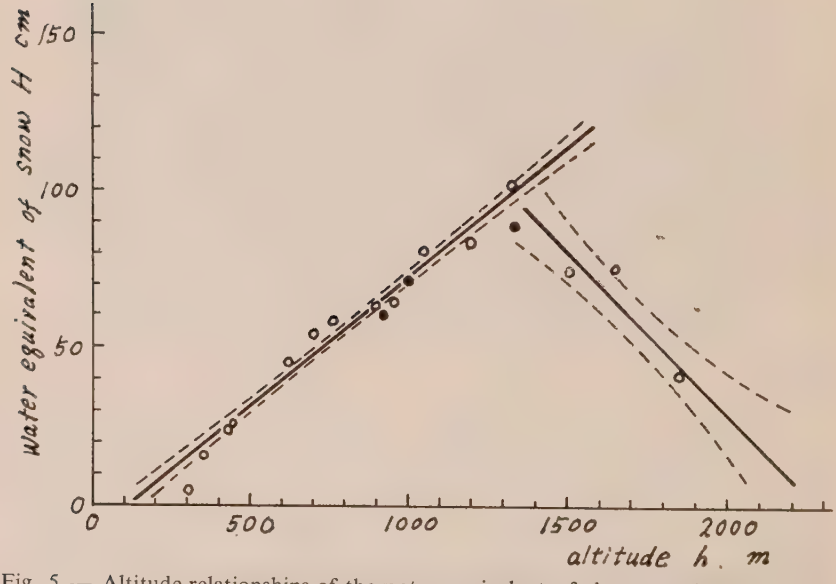

Fig. 5 — Altitude relationships of the water equivalent of the snow deposit in the Chubetsu River Basin in 1953.

TABLE III

Calculation of water equivalent of snow deposit of each zone and total amount

zone	$h_i \pm 100$ m	A_i km²	H_i cm	W $10^7 m^3$	R and corrected value of M	S $10^7 m^3$
above the timber line	above 2000	7.1	25	0.18		0.166
	2000 — 1800	19.3	54	1.22	100%	0.245
	1800 — 1600	32.6	83	2.72		0.241
	1600 — 1400	21.3	112	2.39		0.243
	sum	80.8	—	6.51	M 6.51	0.895
high altitude forest	1600 — 1400	12.7	138	1.76	100%	0.053
	1400 — 1200	10.0	119	1.19		0.030
	sum	22.7	—	2.95	M 2.95	0.083
middle altitude forest	1400 — 1200	15.5	119	1.85		0.040
	1200 — 1000	31.4	100	3.14	84%	0.061
	1000 — 800	32.1	80	2.57		0.046
	800 — 600	16.3	62	0.99		0.022
	sum	95.3		8.55	M 7.18	0.169
low altitude forest	1000 — 800	7.7	79	0.61		0.011
	800 — 600	12.6	62	0.78	87%	0.018
	600 — 400	22.4	43	0.97		0.039
	below 400	0.7	24	0.02		0.002
	sum	43.4		2.40	M 2.09	0.070
cultivated	hilly part	4.2	43	0.18	100%	0.039
	flat part	10.1	24	0.24		
	sum	14.3		0.42	M 0.42	0.039
total		256.0			M 19.15	1.256

h_i : average altitude of a band
A_i : area of a band which has average altitude *hi*
H_i : water equivalent of snow at *h* by eqs. (6) and (7)
W : water equivalent of snow deposit in a band area
R : correction coefficient
S : standard deviation

The last column of the table shows the standard deviation for the estimate of water equivalent of snow deposit for each band. This is obtained by multiplying the standard deviation of H_i at the average altitude of a band by the area of that band. The standard error for the total value is the sum of these values for each band and is shown at the bottom of this column.

Consequently, the total water equivalent of snow deposit in the Chubetsu River Basin at the beginning of April in 1952 was estimated as $(19.15 \pm 1.26) \times 10^7 m^3$.

Total runoff of the Chubetsu River from April 5 to June 30 of this year was estimated as $(25.0 \pm 1.7) \times 10^7 m^3$. The standard deviation of this value is due to the use of the empirical formula for the estimate of runoff of the river from the discharage data of the Eoroshi hydroelectric power plant.

Measurements of precipitation during the thawing season by means of integrating rain gauges at several points in the basin made it possible to estimate the total precipitation over the entire area in this season. The value of this estimate was $12.5 \times 10^7 m^3$.

Runoff coefficient for the water of melted snow was assumed 100% in Sugaya's case in 1948. This is true only when we compare the water esquivalent of snow deposit with the runoff at the end of the snow thawing season, namely at the end of July But in our case in 1952. it was necessary to finish the program at the end of June. According to Sugaya's observation of remaining snow deposit in this basin in the thawing season, the ratio of melted snow to the total at the end of June is known as 80%. Therefore, to calculate the total water equivalent of snow M from the eq. (5), we must take α equal to 80%.

Substituting numerical values obtained above in Q and P and taking $\alpha = 80\%$ and $\beta = 80\%$, M is calculated as $18.75 \times 10^7 m^3$. The water equivalent of snow deposit obtained by the snow survey $(19.15 \pm 1.26) \times 10^7 m^3$ coincides with this calculated value within the range of standard deviation. Therefore, we can conclude that the «Yukomanbetsu survey route» had good representativeness for determination of the altitude relationship of snow deposit in the entire basin.

In 1953 another survey was carried out in this basin in the same manner as in 1952. Altitude relationships of water equivalent of snow deposit along this route are shown in Fig. 5. Inclination of the regression lines is less than those of 1952, and accordingly total water equivalent of snow in the entire basin was less than that of 1952. An estimate of this amount calculated by the method of Table III was $(14.8 \pm 1.2) \times 10^7 m^3$

The water equivalent of snow deposit calculated reversely from the runoff in this year was $(17.2 \pm 2.0) \times 10^7 m^3$. Two values above coincide in the range of standard error, though the coincidence is not as good as in the preceding year.

This year, a supplementary survey route was taken on a mountain side which leads to Mt. Kawun. Locations of snow boring along this route are shown in black dots in Fig. 1. The water equivalent vs. altitude relation at each point is also shown in black dots in Fig. 5. They all fit fairly well to the regression lines of the Yukomanbetsu survey route. This fact seems to be a further proof of the representativemeness of the Yukomanbetsu survey route for the entire basin.

In 1954, the final snow furvey was carried out. But changes on river bed conditions near the intake of Eoroshi hydroelectric plant due to a flood in the preceding year made it difficult to apply the equation for estimating runoff used before. For this reason definite results are not yet avairable from the data of this year.

Discussion

The practical snow survey method is now established after four years of careful testing in the Chubetsu River Basin. Choice of Yukomanbetsu toute proved suitable

for a determination of the altitude relationship of water equivalent of snow deposit for the entire basin. Good representativeness of this route can be attributed to the fact that it is parallel to the main direction of valleys in the basin and it has a gently increasing slope with no abrupt changes.

It will be interesting to compare the results of four years' snow survey. A comparison of total amounts of equivalent of snow deposit M, precipitation in the thawing season P and runoff Q is given in Table IV.

Special attention should be called to the difference of M in 1952 and 1953. In these two years preliminary surveys were carried out at the beginning of February. At that time, snow layers were marked by means of a dye scattered over the snow surface.

TABLE IV

Comparison of the results of snow survey by the practical mathod

year	M 10^7 m^3 measured	M 10^7 m^3 calculated	P 10^7 m^3	Q 10^7 m^3	surveyor
1949	14.8	11.6	11.0	20.8	Chiba
1952	19.2	18.8	12.5	25.0	Higashi
1953	14.8	17.2	11.1	23.5	Higuchi
1954	18.2	—	—	—	Itagaki

These marks made it possible to distinguish the snowfall before and after that time. Fig. 6 shows the altitude relationships of water equivalent of snowfall before and after the 1 st of February in each year. White dots are for 1952 and the black ones are for 1953. We can see that the amount of snowfall after Feb. 1 is almost the same for each year, while the amount before Feb. 1 is varies greatly. Such differences were observed

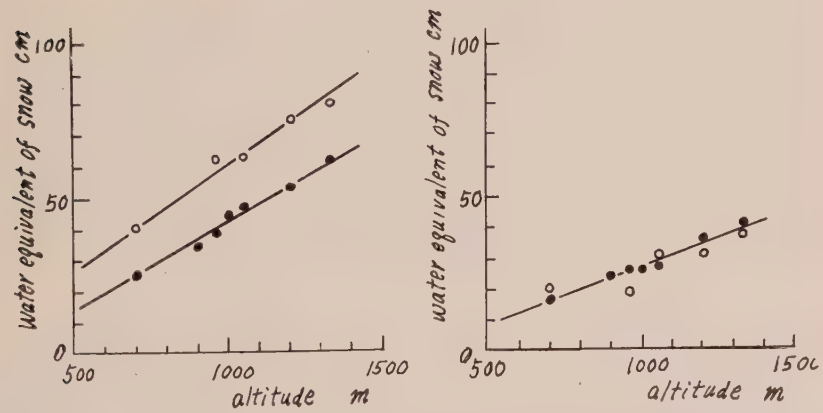

Fig. 6 — Comparison of the snowfall in the Chubetsu River Basin before and after Feb. 1 of 1952 and 1953. (*a*) before Feb., (*b*) after Feb.
● 1952 ○ 1953

widely all over western part of Hokkaido, where the snowfall in winter is mainly brought by the approach of depression from the Japan Sea. As the Chubetsu River flows toward west, it is quite possible that the tendency of snow deposit in this basin coincides with that of western Hokkaido and its magnitude is governed by the snowfall before Feb., during which time the effect of depression from the Japan Sea on the snowfall predominates.

3. The snow survey in the Lake Shikaribetsu Basin

General comments

Lake Shikaribetsu has been used as a reservoir for the Shikaribetsu hydroelectric system of power plants which was constructed in 1953. As this system is to be operated only to make up for the decline in water power for other hydroelectric systems in Hokkaido during winter months, the surface level of the lake falls remarkably during this season. This level must be brought back to the normal by the runoff into the lake before the summer sightseeing season.

Therefore, the amount of water available for use in winter or the maximum water level lowering depends on how much runoff can be expected during the time between the halting power plant operation and the time limit of recovery. The runoff in this period consists of two parts, namely the water of melted snow deposit in the basin surrounding the lake and the water from precipitation.

The first part is known at the beginning of the snow melting season, when we estimate the total water equivalent of snow deposit in the entire basin by a snow survey. The second part requires a forecast of precipitation during a lengthy period, which is still a very difficult meteorological problem. For that reason, ideal operation of the system should be based on the minimum expected snow deposit and precipitation in the thawing season.

The snow survey program which has been carried under a contract of Hokkaido Electric Power Company Ltd. since 1954 aims to ascertain the contribution of melted snow to the runoff and to determine of its variation from year to year. Accordingly, it was proposed to extend the program for several years.

Besides our interest in applying the snow survey method for such a practical purpose, there is another reason which motivates us to carry out this program. The fact that this basin has only 44.8 km^2 area which makes it easy to evaluate total water equivalent of snow deposit and precipitations with considerable accuracy. On the other hand, the Tomabetsu River which was a water outlet before is completely dammed up and has been replaced by the pressure tunnel to the power plants. This makes it possible to estimate the runoff very acculately, provided the water level of the lake and the discharge through the tunnel are known. These are very favorable conditions for a study of the precipitation—runoff relationship in the snow deposit area which will be one of the most interesting subject in hydrology.

The snow survey in 1954

The first snow survey[7]) in this basin was carried out at the end of March 1954. The quantitative method was employed in this year to determine the character of snow deposit distribution in the basin. General basin topography and the locations of the snow borings (white circle) are shown in Fig. 7.

Borings were located in such a way as to cover various altitudes and areal distributions in the basin. In such a «soup - dish» shaped basin, it is quite possible to encounter areal differences of snow deposit corresponding to the main direction

of seasonal wind. These points were selected at open areas in the forest, though it was sometimes difficult to find a favorable location. A density of boring points of 23 per 44.8 km^2 was comparable to that obtained in the Chubetsu River Basin in 1948.

Results of the surfvey are shown in Fig. 8 as the relationships between the water equivalent of snow deposit and altitude. In this figure, black dots represent the values in the northern part of the basin and white ones in the southern part. The boundary between the two zones is shown by a broken line on Fig. 7.

It is clear that these two zones have different tendencies of snow deposition. The water equivalent of snow increases more rapidly with altitude in the northern zone. Regression equations for both areas are expressed in linear form:

$$H = -8.8 + 0.058h \qquad \text{north} \qquad (8)$$

$$H = -1.0 + 0.040h \qquad \text{south} \qquad (9)$$

Fig. 7 — General topography of the Lake Skikaribetsu Basin

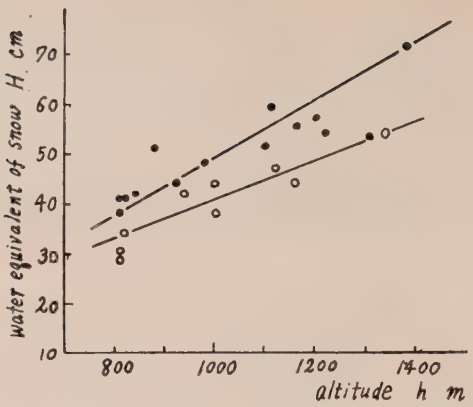

Fig. 8 — Altitude relationships of the water equivalent of the snow deposit in the Lake Skikaribetsu Basin in 1954.

Estimation of total water equivalent of snow in the entire basin was carried out as in the preceding section of this report. The division of bands was much simpler than the case of the Chubetsu River Basin, because the entire basin is below the timber line and altitude ranges from 800 m at lake level up to highest peak of 1470 m.

TABLE V

Calculation of water equivalent of snow deposit in zones and total amount

zone		division of altitude and average altitude *h*	A_i km²	H_i cm	W 10^5 m³
south		800 — 900 (850)	4.0	35	14.0
		900 — 1100 (1000)	4.2	41	17.2
		above 1100 (1200)	0.8	49	3.9
	sum		9.0	—	35.1
north		800 — 900 (850)	8.8	40	35.2
		900 — 1100 (1000)	16.9	49	82.8
		1100 — 1300 (1200)	9.0	61	54.9
		above 1300	1.1	66	7.3
	sum		35.8	—	180.2
total			44.8		215.3
					× 0.85
corrected total					183.0

The procedure for estimation is shown in Table V. The water equivalent H of snow deposit at the average altitude of each band h in this table was calculated by equations (8) and (9) for the northern and southern zones respectively. Correction for the forest cover was applied to the total value of water equivalent of snow deposit, taking 85% as the average correction coefficient. This correction coefficient was evaluated in Table I using an approximate mean density of forest cover from a forest map of the National Forestry Bureau.

The total amount of the water equivalent of snow deposit in the Lake Shikaribetsu Basin at the end of March in 1954 was estimated as $18.3 \times 10^6\ m^3$. The contribution of the northern zone was remarkable.

To determine the contribution of snow deposit to the runoff into the lake, it is necessary to know the total precipitation over the basin during the thawing season. For this purpose seven accumulating rain gauges were placed in the basin. Gauge locations are shown by double circles in Fig. 7. Total amount of precipitation over the basin during the period from April to June was calculated by the weight method of Thiessen using the amount of precipitation obtained by 7 rain gauges above. It was evaluated as $13.3 \times 10^6 m^3$. Through statistical study[8]) of precipitation in the basin area, we have sufficient proof that this amount is far below the normal average value. Accordingly, the contribution of the snow deposit may be considered to be just comparable to the precipitation in this season.

The runoff into the lake was calculated from the data of the discharge to the power plants which were operated until April 20 and from the final level of the lake. Calculated runoff from April to June was $13.7 \times 10^6 m^3$. This amount is so surprisingly small because it gives only 43% for the overall runoff coefficient. The overall runoff coefficient means that the coefficient was calculated for the gross amount of snow edposit and precipitation without any separation as shown in Eq. (5). A runoff coefficient 43% is remarkably smaller than 80% for the one of the Chubetsu River.

The prediction of the water level restoration was made at the end of April, based on the total water equivalent of snow deposit and the statistical estimation of the precipitation in the thawing season which is somewhat indeterminate. Assuming 70% for the runoff coefficient, it was predicted that the lowereing of about 6 m at that time would be restored by the end of July. Actually, this did not occur until August. This discrepancy may be attributed to the small actual coefficient of runoff and to the abnormally low precipitation during the thawing season of this year.

Later activities

The snow survey in this basin has been resumed every winter since 1954. In 1955 several snow borings were added to the ones of the preceding year. They are shown as black dots in Fig. 7. These points provide information about snow deposits in the center of the west side of the basin.

Some effort was made to explain the reason for the small runoff coefficient obtained in this basin. Precise measurements for the comparison of the snow deposits in and out of the forest in the basin were carried out in the spring of 1956. They made it clear that our estimated correction oefficient for forest cover was excessive, specifically, the former value of 85% should have been 65%. When is this value used, the resulting runoff coefficient is 50% as in the survey of 1954.

Another possible reason for this low runoff coefficient is the leakage through some underground passage to the Tomabetsu River. This results in an underestimation of the runoff calculated from the rise of the lake water level. Existence of underground cavities reported at the construction of the pressure tunnel may be favorable to this possibility.

With the exception of brief reports for the operation of the power plants system,

the data of later surfveys are not yet ready for publication. In a few years a more accurate picture of the hydrologic cycle in this basin will be available as a result of these continuing studies.

4. A PRACTICAL USE OF THE RADIOACTIVE SNOW GAUGE.

General comments

It is sometimes very desirable to have a continuous indication of the snow accumulation during the winter in a particular basin. For instance, the use of a water reservoir like Lake Shikaribetsu should be based on the reports of snow accumulation from time to time during the winter. Apart from such an application, it is still an interesting and difficult problem in hydrology to get reliable data of snow accumulation in mountainous region.

For such purposes, the measuring apparatus should have stability of operation for a rather long period of time without any adjustment and be capable of the telemetering data. The radioactive snow gauge invented by R.W. Gerdel and his colleagues[9]) seems to be very useful. This gauge has the advanatge that it indicates the water equivalent of snow deposit directly. The instrument is evidently of practical value in the field of snow hydrology.

We[10]) have been trying to put this type of gauge to practical use in the Lake Shikaribetsu Basin with the intention of substituting this method for the conventional snow survey in the future.

Principle of measurement

The absorption of γ rays by a substance is governed by the following formula:

$$I = I_0 \exp [- k.\rho\, d] \tag{10}$$

where I_0 is the intensity of radiation of the γ ray source, I is the intensity after transmission through a substance of thickness d and density ρ. A quantity called the mass absorption coefficient k is known to be a constant for a given material and independent of the state of that material. This equation shows that the intensity of transmitted γ rays decreases exponentially with the mass of the transmitting material. In the case of snow, this mass is the water equivalent of snow deposit H. Accordingly, the absorption by the snow is expressed by

$$I = I_0 \exp [- k\ H] \tag{11}$$

and is independent of the depth, density or water content of the snow. This is a point which is convenient for the measurement of the water esquivalent of snow. Equation (11) is rewritten as:

$$H = I/k\ (ln\ I_0 - ln\ I) \tag{12}$$

A calibration curve of the relation between I and H makes it possible to determine H by measuring of I.

Construction of the gauge

General construction of the gauge is as follows. A G-M counter is buried just under the ground surface, and a radioactive sourvce of Co_{60} with strong collimation is suspended at a suitable height directly above the counter. This height should be such that the setup does not disturb the natural state of snow dpeosit at this site. General arrangement of the system is shown in Fig. 9.

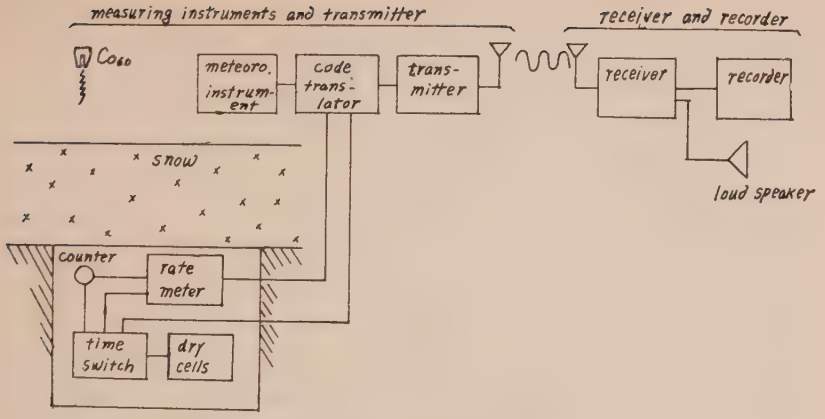

Fig. 9 — Arrangement of the radioactive snow gauge.

The disposition of the source and the counter in the figure is just the opposite of Gerdel's case. This is designed for the purpose of avoiding the change of characteristics of the G-M counters and the lowering of dry cell strength under low temperature. Under this disposition, all temperature sensitive parts are kept at nearly constant temperature of about 0°C. Sprecial G-M counters were made using ethyl-formate for the quenching gas. These tubes have good characteristics even at-35°C.

The output current of the rate meter converted from the pulse counted by the counter tube is used to operate a Morse code system in a radio transmitter. The Morse code system is more convenient than the frequency modulation system used in the United States[11]), because the counting signal is not affected by the noise and it facilitates the transmission of such meteorological data as the temperature in addition to radioactive count.

A time switch which is connected to an electric clock operated by a dry cell was installed in a box underground and it moves the telemetering system twice a day, for several minutes each time.

The most important and critical part of this setup is the rate meter and the Morse code translator. The output current of the rate meter is converted to the deflection of a needle of a current meter. A contact point on this needle slides on a Morse code drum as shown at the bottom of Fig. 10. The exponential relationship as shown in eq. (12) gives a relationships between the movement of the needle and the water equivalent of snow as shown in Fig. 10 (*a*). This makes accurate measurement at higher values impossible.

This inconvenience was eliminated by the use of a remote cutoff electronic tube with the rate meter which makes the output current proportional to $-ln\ I$. The new relationship between the movement of the needle and the water equivalent is approximately linear as shown in Fig. 10 (*b*). Magnitude of error by the Morse code in this case is about 1 cm.

Results of tests

The first gauge was completed at the end of March 1955. It was tested in Sapporo and give sufficiently successful results in operation. In the winter of 1955-56 it was set at a point in the northern part of Lake Shikaribetsu Basin at which point water

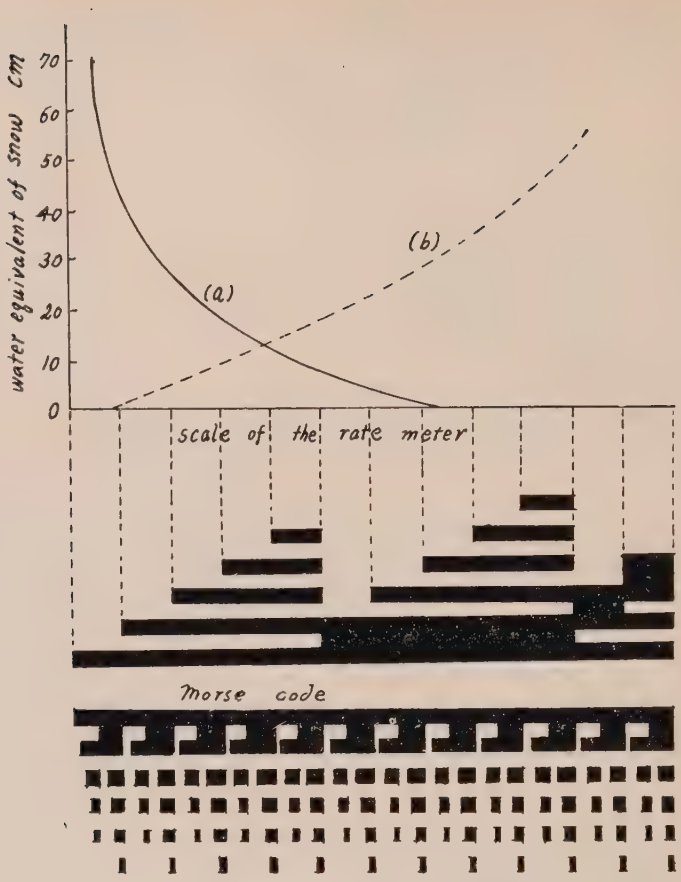

Fig. 10 — Development of the rate meter and Morse code drum assembly.

equivalent of snow deposit shows approxemately average the value for this basin. Some technical difficulty, such as a failure of suspension of the radioactive source, clock stoppage, etc., made continuous telemetering impossible.

On the other hand, many improvements like that of the rate meter assembly described above and employment of a new counter tube were introduced. The second gauge was set with many precautions at the asme place last year and is still under test.

5. Summary and acknowledgement

Snow survey works carried out in the Chubetsu River Basin in Hokkaido were described with a review of the original results of the quantitative method. A practical method in this basin is established by taking a snow survey route whose representativeness was proved by the survey works for four years under varying conditions.

The quantitative snow survey method was extended to the Lake Shikaribetsu

Basin, where the results are directly applied for the use of this lake as a reservoir. The abnormally low runoff coefficient obtained in the thawing season in this basin demands a more precise study of the hydrologic cycle throughout the year.

Comparison of the contribution of snow deposit to the runoff in the thawing season about the Chubetsu River Basin and the Lake Shikaribetsu Basin will be interesting. The ratio of the total equivalent of snow deposit against the total runoff exceeds 60% in the former (see Table IV), while it does not seem to reach to 50% in the latter. Main reason of this difference is the less snowfall in the Lake Shikaribetsu Basin. The effect of the mountain barrier on the snowfall in Hokkaido which generally accompanies notrh-west seasonal wind is remarkably revealed in this comparison.

The use of the radioactive snow gauge in the Hokkaido area was briefly described and technical improvements achieved were mentioned.

The snow survey works reported here were carried out under Prof. Nakaya's kind and encouraging guidance. The author wishes to express his cordial gratitude to Dr. Sugaya for his kind advice and help for the use of the equipments. Special mention is likewise made of Messrs K. Higuchi and K. Itagaki for their cooperation in carrying out the works and preparing this paper.

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STUDIES OF THE FREQUENCY OF SNOWFALL IN ENGLAND, 1668-1956, AND THE RELATIONSHIP WITH GLACIER BEHAVIOUR

G. MANLEY

Abstract

In London and its neighbourhood, observations of the occurrence of snow and sleet have been kept by various individuals and institutions almost continuously since 1668. While these observations differ widely in quality it has been possible to compile a sufficiently reliable assemblage of all days on which snow has been observed to fall since 1811. For the earlier years standards of observations are less good, but by making use of suitable reduction factors the records can be reasonably equated with those kept under more modern conditions.

The results show (1) fluctuations, but no serious overall change in the likelihood of snowfall; (2) a broad relationship with the mean temperature of the six colder months. (3) The curve showing the frequency of snowfall on given days does not follow a uniform course, but displays irregularities, some of which are very persistent and agree with the «singularities», or irregularities in the annual march of temperature, shown by other European records. (4) The fluctuations in the frequency of snow at London are broadly related to those in Scotland, and may thus provide a rough index of the mountain accumulation; but they show little agreement with the behaviour of glaciers. The latter is more closely associated with the fluctuations in the accumulated temperature of the summer months in England.

For several years the relationship between the atmospheric circulation over the North Atlantic and the behaviour of glaciers in the adjacent countries has been a subject of discussion. In lowland countries such as England where there are no glaciers, past variations in the frequency of snow and snow-cover should be put on record and are presented here, in virtue of the exceptional length of the period of observation.

In the European lowlands west of the Rhine winters are prevailingly mild. Hence the occurrence of snow or sleet, whether it lies or not, is a matter of general comment. For the London area, before the days of official meteorology, we have journals kept by numerous observers which permit the compilation of a nearly complete record of the frequency of snowfall back to 1668.

Observations of the number of days with snow or sleet are however of unequal value. In England observers vary widely in their readiness to make note of a few flakes of melting snow falling in chilly rain; or, a few drifting flakes of dry powdery snow in cold weather. Scrutiny and comparison of older journals, as well as those kept at the present day, shows that a single observer can normally pay little attention to what happens in the night.

S.E. Ashmore (Q.J. Roy. Met. S. **78**, 1952, p. 629) has discussed the probable differences in the annual total of days deriving from present-day climatological stations of several types: he recognises five grades or qualities of observation (A, A-, B,C,D) and provides approximate reduction factors. His findings apply well to the numerous stations in the London area. Grade A is characteristic of such stations as Croydon or London Airport, where a constant day and night watch is kept by professional observers. Before the days of official meteorology and precise definition of terms, it is still evident from the nature of the entries in journals that the keenness

of some amateur observers was such that we can accept their observations as approximating to A standards. In general, however, most of them can be graded C, which broadly corresponds with the standard of a normal third-order climatological station to-day, i.e. a station at which instruments are read once daily.

A scrutiny of the observations in the London area enables a statement to be made of the number of days with snow or sleet observed to fall, by A standards, from 1811 onward. These observations are related to an altitude of 50-60 metres above sea level (150-200 feet). The principal component, covering 1841 to 1950, is provided by the Greenwich Observatory (49 m.); but the daily records from other places within the London basin have also been examined and assembled. In general the records from three such stations, each of first-class or nearly first-class quality, have been used, covering a triangle with sides not exceeding 25 km. and of negligible relief. To use three stations is advantageous, as it enables one to recognise and include a number of days with scattered showers which might be observed at one station and not at the others. A few reports in respect of which uncertainty of definition exists were checked against the temperatures of the days in question. The earliest and latest dates with snowflakes reliably observed were 25 September, 1885 and 27 May, 1821.

The principal elements of the resultant assemblage of observations are shown in the diagrams below, for these I have to thank Miss E.M. Shaw, M. Sc.

London and district: observations before 1811

These are all to be described as «amateur», and derive from men of varying professions, character and alertness. No precise definition of what was to be observed had yet been formulated; and the majority refer to low-lying stations near the level of the river. The averages for 1931-1955 at Kew (16.5, at 5 m.) and Croydon (19.0, at 66 m.) incidate the need to allow for an appreciable effect of altitude.

During the greater part of the 18th century two stations are available, assuming that use can fairly be made of lowland records within a radius of 40 km., for which there is meteorological justification. In general they appear to have been quite well kept in the daytime hours, and for the most part can be ranked «C» by present-day standards. For the years 1786-1807 there is also a record kept by an instrument maker in the City of London which can undoubtedly be considered «B», both from the nature of the entries and in comparison with other ocntemporary records kept not only near London but also in the Midlands and North. The second diagram (Fig. 2) below will show not only the accordance of trend but also how far the totals by «A», «B» and «C» standards can be expected to differ.

Allowing for the conditions of observation, nearly all the available observations beginning with the winter 1668-69 up to that of 1810-11 have been regarded as of «C» standard and assembled below. Two qualifications must be made: first, for the years 1713 to 1727 the only available set of daily observations must be regarded as of «D» standard, if not even lower, and for those years estimates have been compiled using Ashmore's reduction factor to bring the totals into comparison with those of later years. Secondly, for the years 1690-91, 1693-96, and 1707-12 we have no adequate daily observations at all. For those years only the roughest estimates can be made, based on very general descriptions of the weather of those years.

The results, displayed below (Fig. 1), indicate no serious change in the overall incidence during the past 300 years. When the annual totals are assembled in the form of decadal running means, a fairly consistent relationship with the mean temperature of the six colder months (November-April) can be recognised, (Fig. 2). There is a noteworthy variation in the incidence of snowfall around given dates during

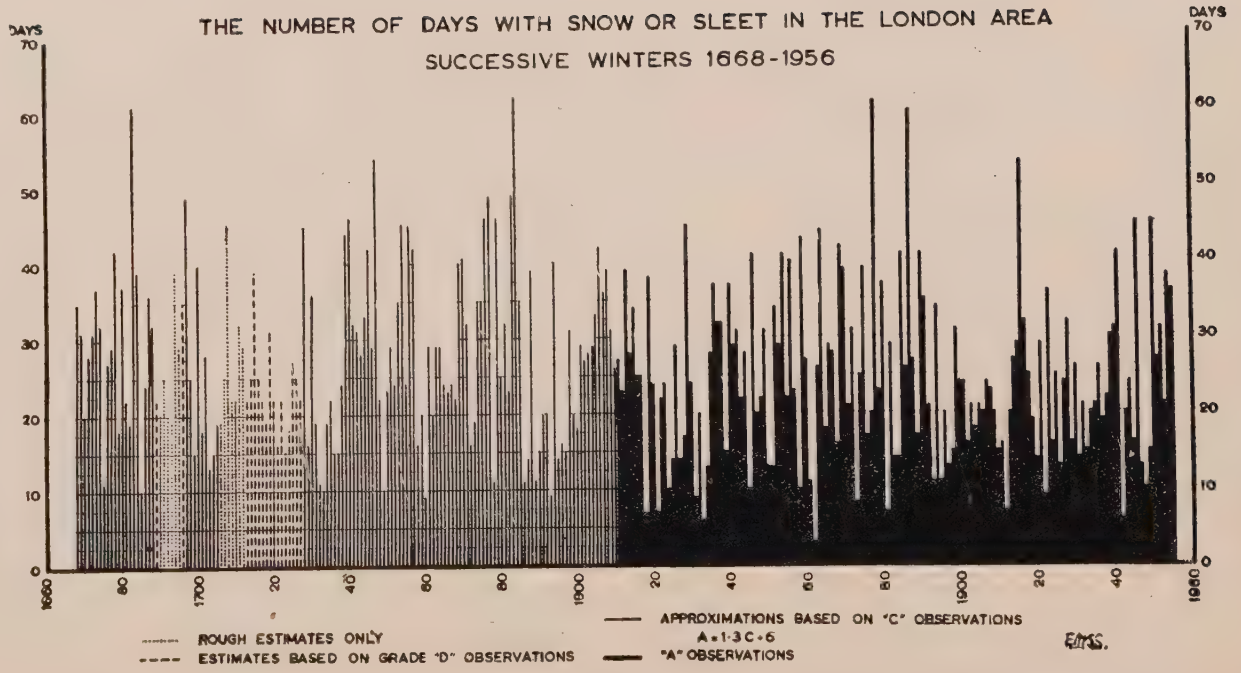
THE NUMBER OF DAYS WITH SNOW OR SLEET IN THE LONDON AREA
SUCCESSIVE WINTERS 1668-1956
DAYS
70
60
50
40
30
20
10
0
1660
80
1700
20
40
60
80
1800
20
40
60
80
1900
20
40
1960
ROUGH ESTIMATES ONLY
ESTIMATES BASED ON GRADE "D" OBSERVATIONS
APPROXIMATIONS BASED ON "C" OBSERVATIONS
A=1·3C+6
"A" OBSERVATIONS
EMS.

Fig. 1

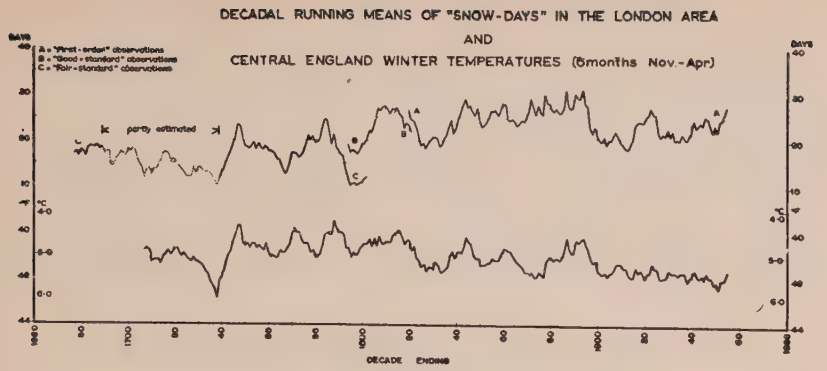
DECADAL RUNNING MEANS OF "SNOW-DAYS" IN THE LONDON AREA
AND
CENTRAL ENGLAND WINTER TEMPERATURES (6months Nov.-Apr.)
DECADE ENDING

Fig. 2

the winter months. The diagram (in Fig. 3) illustrates the average frequency with which snow has been recorded during the «A» observation period (145 years) and the earlier «C» observations (89 years); the totals for each day being smoothed in successive five-day means. which in some measure eliminates chance variations in the totals. That certain peaks *centred* about Nov. 18-23, Dec. 15-16, Jan. 7-11, Feb. 1-4, Feb. 19-20, March 9-12, April 11-14 stand out in each of the records appears to point to tendencies in the west European climate deserving of further study; some of them conform with Buchan's spells, while others may be found acceptable in Central Europe. Some of them agree with irregularities in the annual march of temperature at Oslo (Hesselberg, Climate of Oslo, Norsk Geog, Tidskr. **2**, 1929). The rapid rise in the first half of January is noteworthy. None of the peaks is however so strongly defined as might appear; days around the peaks have sabout 25% greater chance of snow falling than days in the hollows of the curve, which is not a large difference.

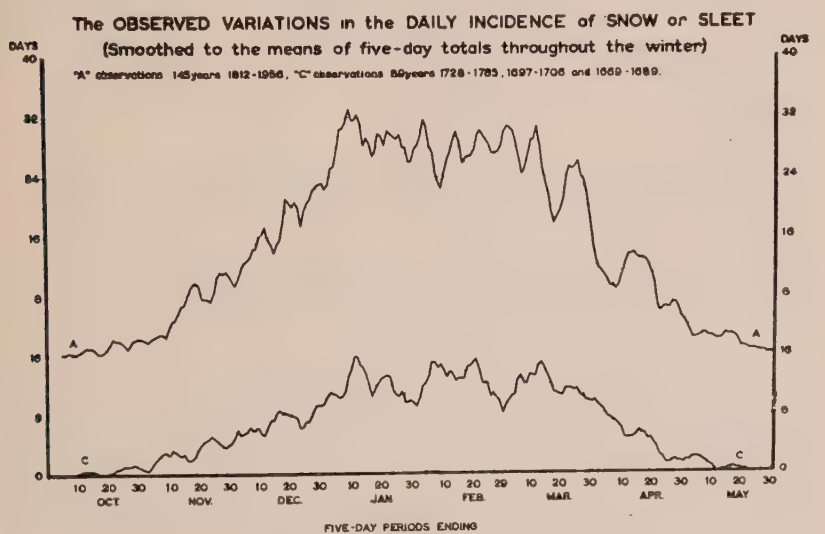

Fig. 3

Observations of the frequency of snowfall are available for various periods and places in Scotland, back to 1770. They are of variable quality, and it is not easy to standardise them on the same principles as are practicable for London, as the records do not always overlap. Nevertheless, assemblage of the data (not shown here), in the form of decadal running means, indicates that the overall trend of the variations in the frequency at London is satisfactorily representative of that of the north.

It may be estimated that, of the days on which snow or sleet is observed to fall at lowland stations with «A» standards, about two-thirds are likely to make a significant contribution to the accumulation at higher altitudes, say above 2000 feet.

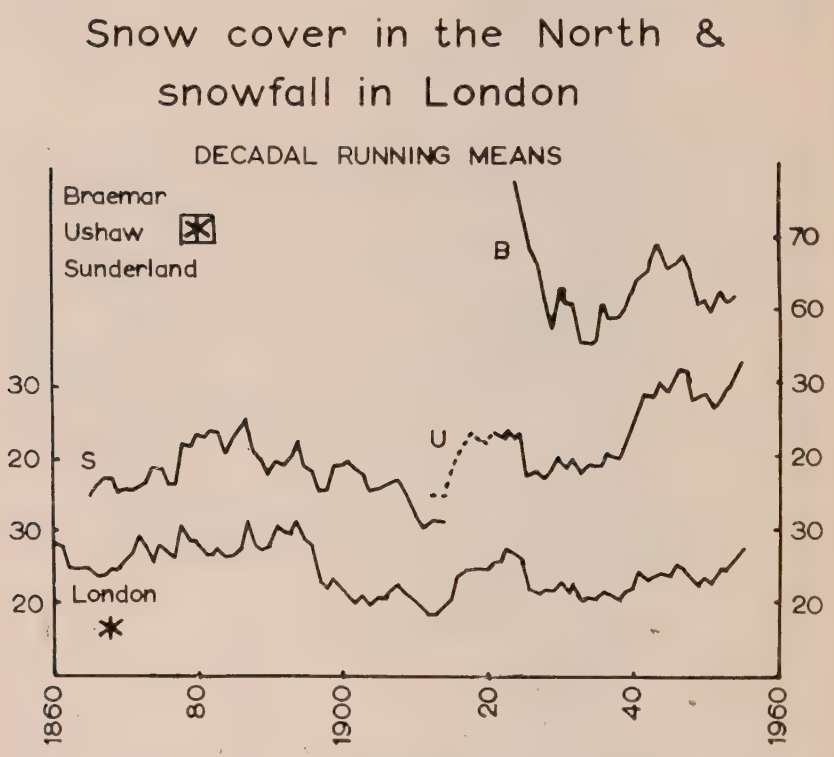

Fig. 4 — Comparative trends of number of days of snow cover in N.E. Scotland and N. E. England, with snow at London.

Hence as the variations in the London totals are reasonably representative of those of coastal districts in the north, they should also be reasonably representative of the variations in the total upland accumulation. We know that the rate of increase with altitude in the duration of upland snow-cover, about 10 days for 65 metres in N. Scotland and 75 m. in N. England, is considerably more rapid than that for Central Norway (10 d. for 100 m.) or the Swiss Alps (10 d. for 100 m. overall). Cf. Moriköfer, U.G.G.I. *Oslo*, 1948, and Manley, Geogr. Ann., 1949.

We lack long-term statistics of the duration of mountain snow-cover, and we have very little reliable statistical information at all with regard to snow-cover, even for lowland stations, before 1912. Such information as we have, when plotted (Fig. 4), does show a fair overall relationship between the vicissitudes of the duration of snow-cover and that of the frequency of snowfall at London. Hence it would appear that the frequency, the consequent accumulation of snowfall, and the persistence, are chiefly a function of the mean temperature of the six colder months, November-April, with regard to the uplands.

Winter temperature is not so closely related to glacier variation. The relationship between the vicissitudes of glaciers in Iceland and the fluctuations of summer temperature in England is more evident (Fig. 5). The dates of the peaks and hollows

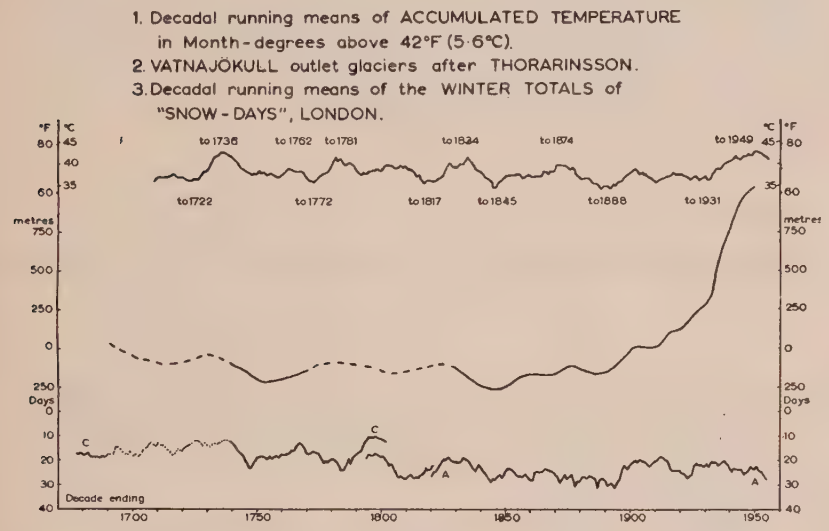

Fig. 5

on this curve accord to some degree with the behaviour of the Alpine glaciers. In contrast, the same diagram shows that there appears to be much less relationship between the glacier behaviour and the winter frequency of snowfall in England.

Studies of the frequency of snowfall therefore suggest that (1) there has been little overall change in the character of the winters during the past 300 years; but that the proportion of severe winters to the total has varied from time to time; and (2) in regard to the development of persistent snow-cover in N.W. Europe and glaciation it is more important to consider in what ways the spring and summer temperature might effectively be lowered. (3) The fact that certain groups of days in the winter and spring appear more likely to give snow than others, and that such tendencies have persisted over nearly 300 years, may draw renewed attention to the problem of singularities in the North-West European climate. Lastly (4) the fact that the period 1668-1710 was not marked by any predominance of adverse seasons other than about 1671-75 and in the later 1690's, supports the view that the glacier readvances within that period were probably not so marked as those of the 19th century; or the Alpine readvances of the early 17th century.

AIR PERMEABILITY OF SNOW

J. A. BENDER
Snow and Ice Basic Research Branch
U. S. Army Snow Ice and Permafrost Research Establishment

Abstract

An apparatus has been constructed for determining very accurately the air permeability of granular materials under low pressure gradients. Screened fractions of snow, glass beads and well rounded soil were systematically investigated over a wide range of air velocities, to find the effect of grain size, porosity, shape and flow velocity on the permeability. The permeability of a great number of natural snows were determined at various densities, and a grain-size analysis made of the natural mixtures. All of the results obtained under laminar flow conditions may be represented by a single relationship $K = \frac{an\,N}{N - n}$, where K is the air permeability, a is a function of grain size, n is the porosity and N is a structure factor. It is believed that a and N are parameters that uniquely define a given snow.

Introduction

Darcy (1856) first stated the basic law for laminar flow through a porous medium as

$$v = k\frac{\Delta P}{L} = ki \tag{1}$$

where v is the velocity of flow, ΔP is the pressure drop across a sample of length L, and k is the permeability coefficient. A great many investigations have been made to understand better the influence of the porous material on k. Excellent reviews on the general flow of fluids through porous materials have been made by Muskat (1946) and by Carman (1956).

Bader (1939) was the first to make a thorough study of the air permeability of snow, and more recently work has been done by Ishida and Shimizu (1955) and Bender (Bader, *et al.*, 1955).

It was shown by Muskat (1946) that the isothermal laminar flow of a gas through a porous medium may be represented by

$$Q = \frac{k}{\eta}\frac{P_2^2\text{-}P_1^2}{2P_{av}} \times \frac{A}{L} \times t \tag{2}$$

which for small pressure difference may be written as

$$v = \frac{k}{\eta}\frac{\Delta P}{L} \tag{3}$$

It is evident that if the viscosity η of the fluid and the area A and length L of the sample are known, it is only necessary to measure the rate of flow and the pressure difference.

Various types of equipment and methods (*see* Wenzel and Fishel, 1942; Carman, 1956) have been developed for the measurement of permeability. In the present work because only small pressure differences were to be used and work had to be performed at temperatures below freezing, it was necessary to develop apparatus and techniques to fit these needs.

The work was initially started in a cold room maintained at —5C (23°F). However, it was found that the permeability of a snow sample increased with time due to the snow metamorphism. This resulted in the decision to do all of the work in a large cold room maintained at —20C ± 1° (—4°F).

A «Trimount Micro-Manometer» with a range of 0 to 150 ± 0.005 mm of water pressure differential was used. A «Dow Corning 200» silicone liquid having a viscosity of only 2 cs at —20C was satisfactory in the micro-manometer. One end of the manometer was always open to the atmosphere.

Commercially available instruments for measuring fthe flow of air, such as wet test gas meters and rotameters, were tried but did not function satisfactorily at the low temperatures. Instead of trying to measure directly the flow of air through the sample, two types of apparatus were constructed to measure the amount of air displaced at a constant rate.

The apparatus used for measuring small rates of flow 0.06 to 0.8 cm^3/sec is illustrated in Figure 1. It is similar to most constant head permeameters except that a liquid was used to displace the air which is forced through the sample. The rate of flow of the liquid was measured by weighing and timing with a stopwatch.

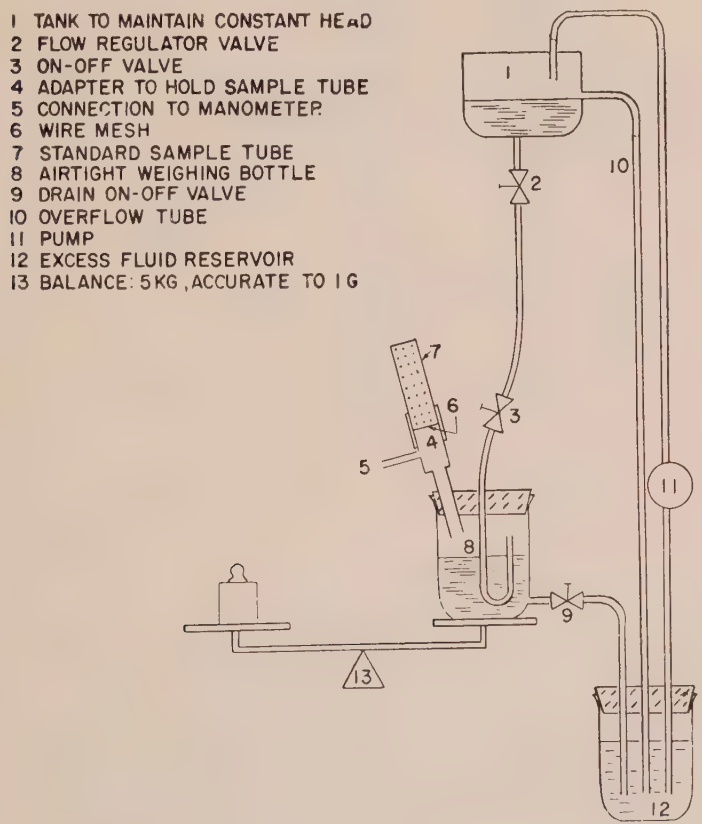
1 TANK TO MAINTAIN CONSTANT HEAD
2 FLOW REGULATOR VALVE
3 ON-OFF VALVE
4 ADAPTER TO HOLD SAMPLE TUBE
5 CONNECTION TO MANOMETER
6 WIRE MESH
7 STANDARD SAMPLE TUBE
8 AIRTIGHT WEIGHING BOTTLE
9 DRAIN ON-OFF VALVE
10 OVERFLOW TUBE
11 PUMP
12 EXCESS FLUID RESERVOIR
13 BALANCE: 5KG, ACCURATE TO 1G

Fig. 1 — Permeameter using liquid to displace air.

In actual practice, the one regulator valve was replaced by a manifold of three valves which were left at different rates of flow and these were checked only at the beginning and end of the operation on a given sample.

For greater rates of flow, 0.5 to over 2000 cm³/sec, the apparatus shown in Figure 2 was used. It consisted of three concentric cylinders. The inner, A, and outer,

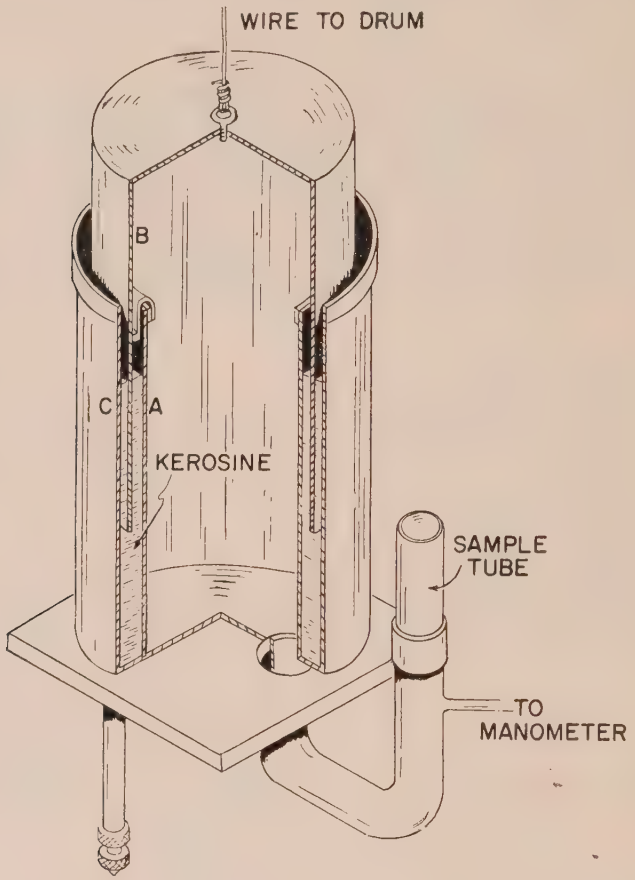

Fig. 2 — Constant velocity permeameter.

C, cylinders were fixed and the space between held kerosine. The middle cylinder, B, could move up and down with the kerosine acting as an air seal. Air could be drawn or forced through the sample by moving the middle cylinder up or down. The movable cylinder was raised or lowered by winding or unwinding a wire on a drum. The rps of the drum could be changed by a series of gears connecting the drum to a synchronous motor.

The sample tube used for almost all of the measurements was the standard SIPRE polished stainless steel snow tube; length, 18.9 cm (7.4 inches), inner diameter 5.8 cm (2.28 inches).

The procedure used for testing snow samples was to fill the standard tube, weigh it to obtain the density, and determine the permeability over a wide range of velocities. The sample then was artificially compacted a few millimeters by means of a hand press. The density was determined by measuring the new length of the compacted sample and the permeability determined again over a wide range of velocities. The above procedure was used to obtain curves for several densities of the same sample. The procedure for glass beads and soil particles was similar except that the tube was tapped to obtain higher densities.

All of the densities were changed into porosities by the relationship

$$n = \frac{\rho t - \rho a}{\rho t} \times 100 \qquad (4)$$

where n is the porosity, ρt the true density of the particles and ρa the apparent density.

Work by Bader (1939), verified in this investigation, showed that the snow sample when compacted was still quite homogeneous and that there was little breaking of individual crystals.

Experimental results and discussion

a. *Range of Darcy's Law*

Each fraction of the screened snow was tested over a wide range of velocities by the methods previously described. Figures 3 and 4 show typical velocity vs pressure gradient curves obtained for two of the fractions. It may readily be seen that there

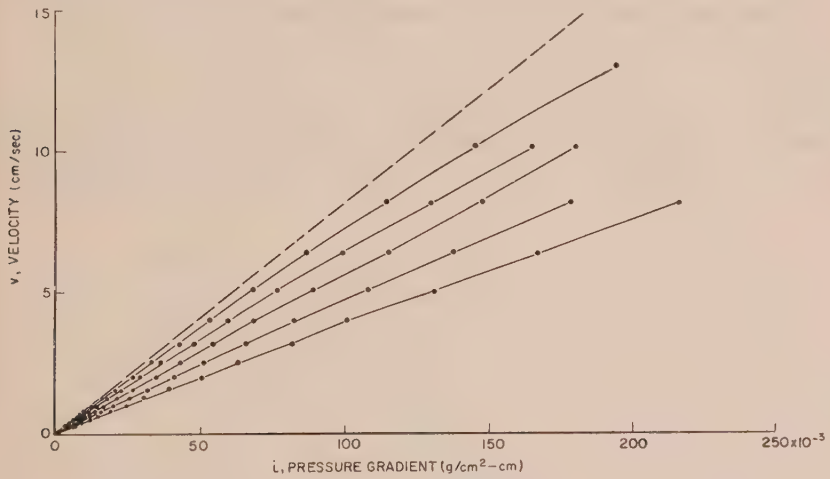

Fig. 3 — Velocity *vs* pressure gradient, —1.190 + 1.000 mm snow fraction at different porosities.

is nowhere a definite break in the curve to show the change from viscous to turbulent flow. This is to be expected as the curves represent the summation of flow through a large number of tortuous channels. It is also seen that the amount of deviation from a linear relationship is not only a function of the velocity and particle size, but of the porosity as well. It is this latter effect which made any attempt to formulate

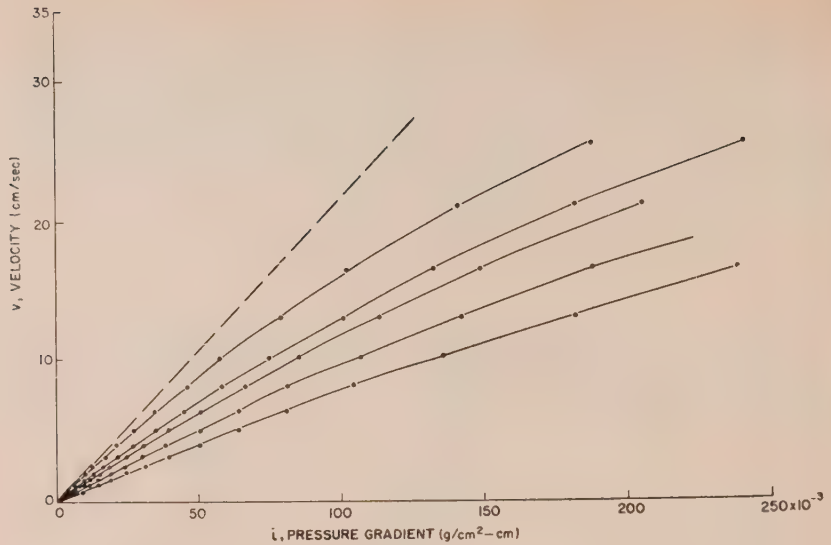

Fig. 4 — Velocity *vs* pressure gradient, —2.000 + 1.680 mm snow fraction at different porosities.

the conditions for the range of viscous flow by a type of Reynold's number quite useless. There is also an additional effect due to particle shape.

Attempts were made to express the curves empirically by

$$i = c_1 v + c_2 v^2 \tag{5}$$

and

$$i = c_3 (v)^r \tag{6}$$

where c_1, c_2 and c_3 were functions of n and particle size and the exponent r varied between 1 and 2. However, none of these equations was successful over the whole

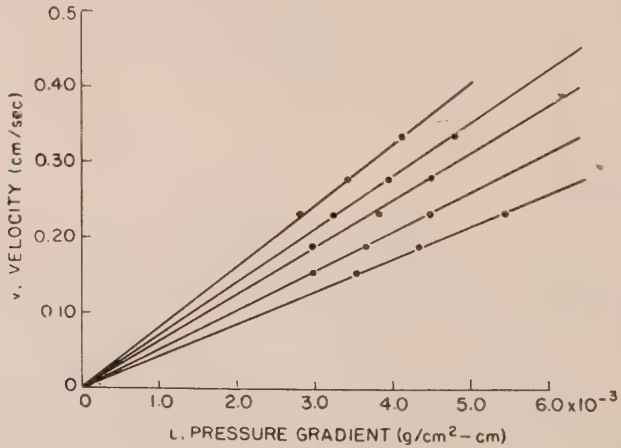

Fig. 5 — Velocity *vs* low pressure gradient, —1.190 + 1.000 mm snow fraction at different porosities.

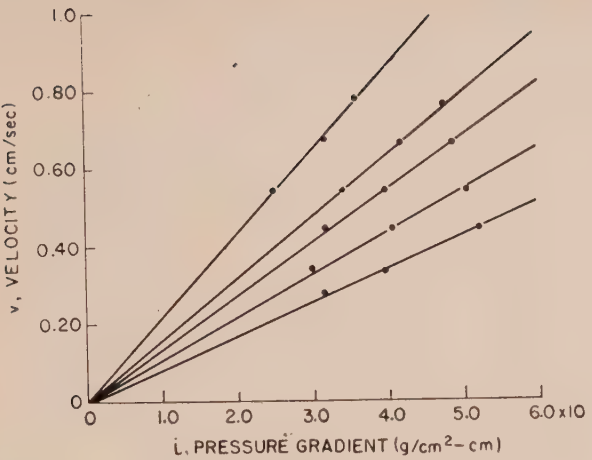

Fig. 6 — Velocity *vs* low pressure gradient, —2.000 + 1.680 mm snow fraction at different porosities.

range of porosities and grain sizes. The results in the viscous range for the same two fractions are shown in Figures 5 and 6.

Bader (1939) recommended flow velocities of no more than 15 cm/sec for fine-grained snow; 6 cm/sec for medium-grained snow; and 2-3 cm/sec for coarse-grained snow. In the experiments on air permeability of natural snows by Ishida and Shimizu (1955), they found the flow was laminar up to 1.5 cm/sec. The author recommends flow velocities no greater than 5 cm/sec for fine-grained snow (less than 0.8 mm diameter); 2 cm/sec for medium-grained snow (0.8 — 1.2 mm diameter) and 1 cm/sec for larger-grained snow .The majority of the tests were made at flow velocities of less than 0.5 cm/sec.

b. *The effect of porosity on the permeability of snow*

The density of snow varies from considerably less than 0.1 for a light fluffy new snow to that of ice. The porosity may readily be computed from the density by the relationship

$$n = \frac{\gamma_i - \gamma_s}{\gamma_i} = 1.00 - 1.09\,\gamma_s \qquad (7)$$

where γ_i = density of ice = 0.917

γ_s = density of snow.

The pores in snow are usually entirely interconnecting, but at high densities the material may no longer be permeable, although still porous. By this difference snow and ice are defined, i.e., snow is permeable, ice is not. Bender (Bader, *et al.*, 1955) found that zero permeability could be expected when the density reached approximately 0.77. Thus there is permeability in a range of porosities from 16% to 89%. Most previous investigators on other media have only been concerned with a few percent change in porosity. Plots of the permeability coefficient K versus the porosity

for a new snow, a large-grained snow and a fine-grained snow are shown in Figures 7, 8 and 9, respectively. K is defined by the equation

$$K = \frac{Q}{At}\frac{\Delta P}{L} = \frac{v}{i}.$$

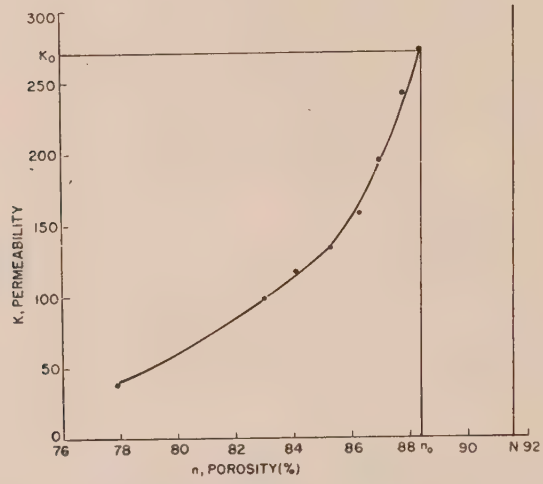

Fig. 7 — Permeability *vs* porosity for a new snow.

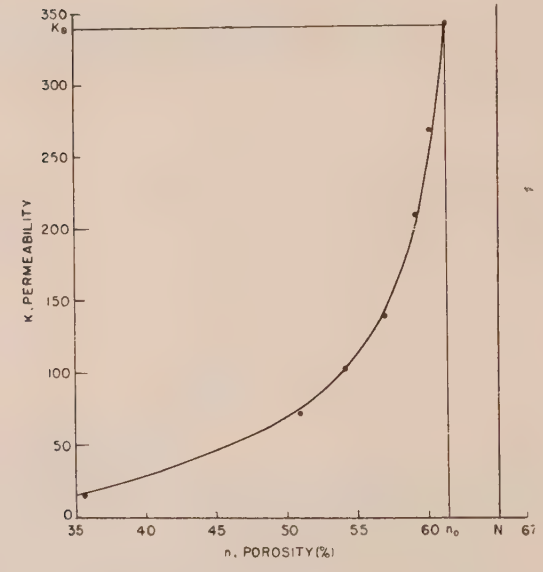

Fig. 8 — Permeability *vs* porosity for a large-grained snow.

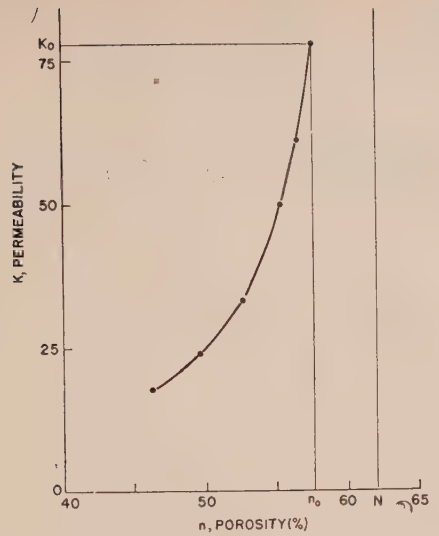

Fig. 9 — Permeability *vs* porosity for a snow over a year old.

Q = volume of air in cm^3
A = cross-section of tube in cm^2
t = time in seconds
P = pressure head in cm of water
L = length of snow sample in cm
n_0 = original porosity before artificial compaction
K_0 = original permeability before artificial compaction

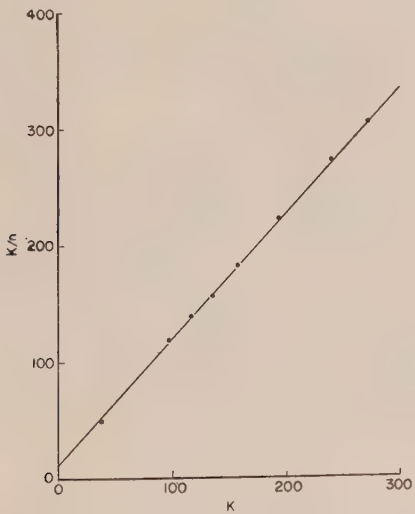

Fig. 10 — K/n *vs* K for a new snow.

Bader (1939) found that his data on permeability could be represented by a straight line when K/n was plotted against K. In this case

$$\frac{K}{n} = a + \frac{K}{N} \text{ or } K = \frac{anN}{N-n}$$

where a is the intercept and N the reciprocal of the slope.

Plots of K *vs* K/n for the three snows shown in Figures 7 - 9 are given in Figures 10 - 12.

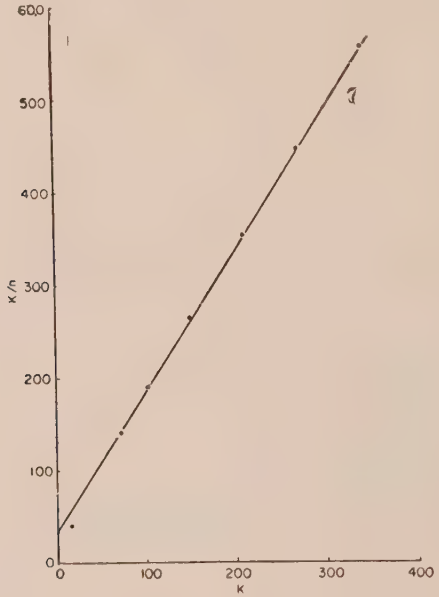

Fig. 11 — K/n *vs* K for a large-grained snow.

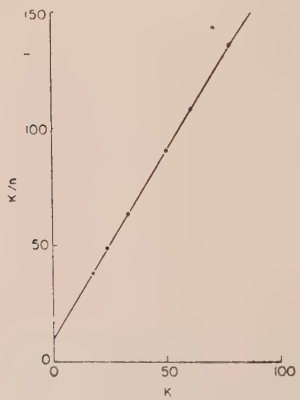

Fig. 12 — K/n *vs* K for a snow over a year old.

A systematic determination of the permeability of natural snows, screened fractions of snow, glass beads, and a round-grained soil was made to investigate the effect of porosity and grain size on the permeability coefficient.

The following functions were investigated for a given natural snow or screen fraction:

$$K_1 = \frac{anN}{N\text{-}n}$$ Bader (1939) empirical equation

$$K_2 = \frac{Bn^2}{(1\text{-}n)^3}$$ Empirical equation

$$K_3 = \frac{Cn^2}{1\text{-}n}$$ Burke and Plummer (1928) theoretical equation

$$K_4 = \frac{Dn^3}{(1\text{-}n)^2}$$ Kozeny (1927) theoretical equation

$$K_5 = \frac{an^2N}{N\text{-}n^2}$$ Empirical equation

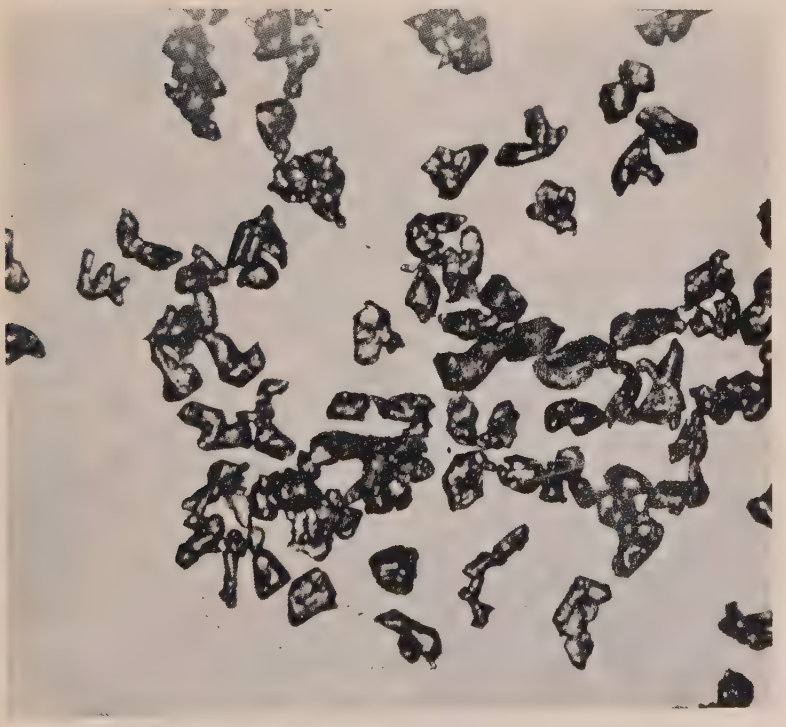

Fig. 13 —0.590 + 0.500 mm snow fraction, 9X.

The values for the constants in each case were determined by the method of least squares. The Gauss criterion (*see* Worthing and Geffner, 1943, page 260),

$$\Omega_i = \frac{\Sigma\,(K_{exp} - K_i)^2}{S - m} = \text{minimum},$$

where K_{exp} is the experimental value; K_i is the computed value, S is the number of observed pairs of data, and *m* the number of arbitrary constants, was used to see which equation best fitted the data. The results are tabulated in Table I.

TABLE I

Material	Ω_1	Ω_2	Ω_3	Ω_{1-}	Ω_1	Range of *n* %	S
Glass Beads	0.02	0.01	0.28	0.05	0.007	31.4—35.0	4
New Snow	128.	150.	1,205.	128.	31.	83.0—88.4	7
Soil	0.04	0.18	1.52	0.25	0.02	49.5—55.1	4
Snow Fraction	0.10	0.08	0.94	0.05	0.04	44.9—51.9	5
B Snow	0.21	48.	181.	54.	0.59	46.2—57.6	6
C Snow	4.8	115.	229.	115.	9.4	42.4—54.6	6

Fig. 14 —0.590 + 0.500 mm glass bead fraction, 9X.

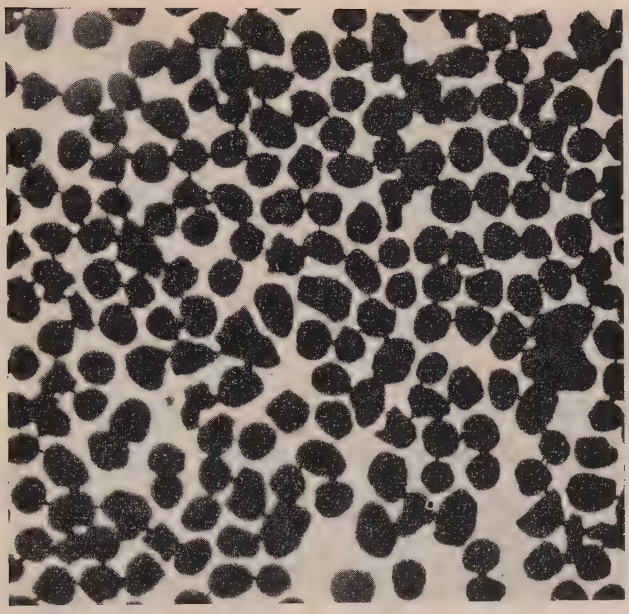

Fig. 15 —0.590 + 0.500 mm soil fraction, 9X.

Fig. 16 —1.000 + 0.840 mm snow fraction, 9X.

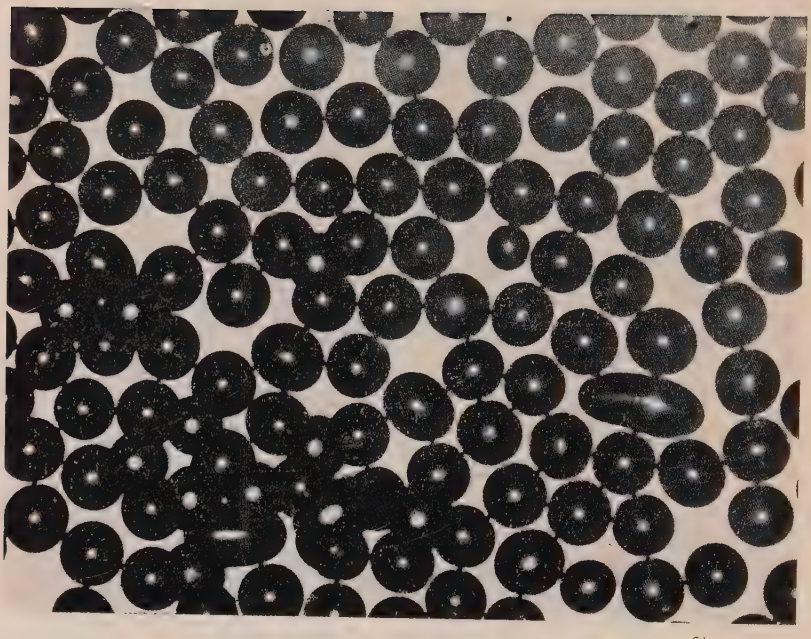

Fig. 17 —1.000 + 0.840 mm glass bead fraction, 9X.

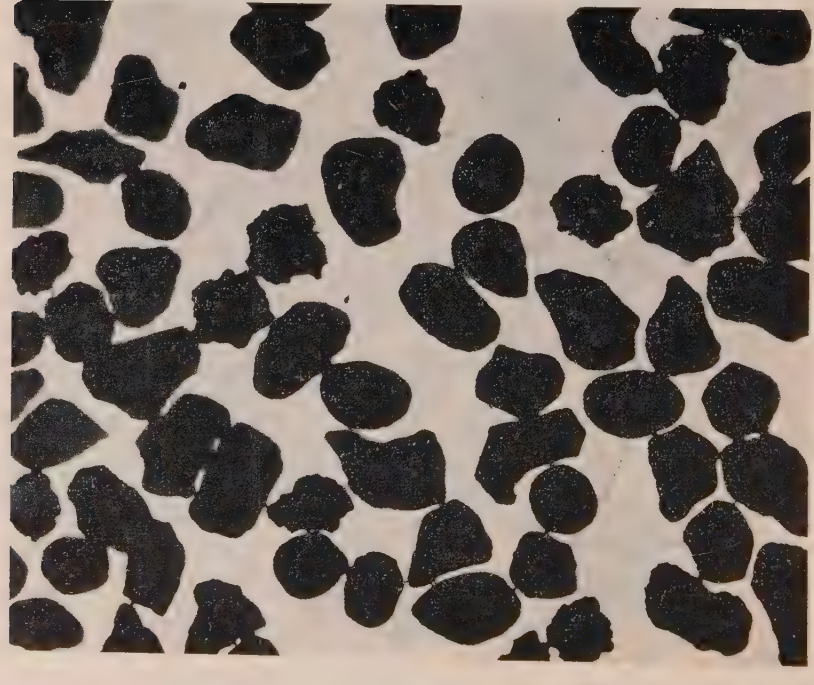

Fig. 18 —1.000 + 0.840 mm soil fraction, 9X.

It may be seen from the table that Bader's function $K_1 = \frac{anN}{N-n}$ or the frunction $K_5 = \frac{a\,n^2N}{N-n^2}$ are the most suitable, especially for snow.

Typical screen fractions of the three types of materials investigated are shown in Figures 13-18. The experimental value for a and N are given in Table II. All values were computed by the method of least squares. In addition to the permeability studies on screened fractions, a number of tests were made on natural Greenland snows. The permeability of a given natural snow was determined for at least five different porosities and a and N computed by means of the Bader equation, using the least squares method. A screen analysis was then made and the mean grain size d calculated from it. The results are given in Table III.

TABLE II

Screened Fractions

Screen size	Screen $\sqrt{m_1 \times m_2}$ size	Snow a	Snow N	Soil a	Soil N	Glass a	Glass N
—0. 149+0.074	0.105			0.14	0.725		
—0. 210+0.149	0.176			0.39	0.676		
—0. 250+0.210	0.229	1.59	0.584	0.66	0.652		
—0. 297+0.250	0.273	2.29	0.606	0.97	0.641		
—0. 350+0.297	0.322	2.59	0.582	1.29	0.630		
—0. 420+0.350	0.383	3.78	0.612	1.99	0.620		
—0. 500+0.420	0.458	4.86	0.623	3.27	0.628	7.22	0.460
—0. 590+0.500	0.543	5.65	0.585	4.46	0.625	11.11	0.466
—0. 710+0.590	0.647	6.33	0.589	7.66	0.618	17.17	0.507
—0. 840+0.710	0.772	8.29	0.645	10.70	0.597		
—1.00 +0.840	0.916	12.11	0.678	18.19	0.613	38.11	0.492
—1.190 +1.000	1.09	16.69	0.622	20.33	0.573	49.49	0.504
—1.410 +1.119	1.29	19.57	0.621	29.55	0.558	58.16	0.479
—1.68 +1.41	1.59	25.40	0.698				
—2.00 +1.68	1.83	27.24	0.656				
—2.38 +2.00	2.18					110.8	0.485
—2.83 +2.38	2.60					169.4	0.502
—3.36 +2.83	3.08					356.9	0.541
—4.00 +3.36	3.67					436.4	0.535

c. *Bader's Equation*

Log-log plots of the values of a in the equation $K = \frac{anN}{N-n}$ against the mean grain size d of natural and screened fractions of snow, fractions of glass beads, and fractions of a soil are shown in figures 19-21, respectively. These show that for snow, $a = 16.8\ d^{1.63}$; for glass beads, $a = 36\ d^{2.0}$; and for the soil, $a = 16.5\ d^{2.17}$.

TABLE III

Natural Snows

Snow	*d*(mean)	*a*	N	N/n^0
B	0.82	10.	0.62	1.076
C	0.74	7.	0.57	1.043
D	0.87	11.	0.59	1.062
E	0.73	11.5	0.58	1.077
F	0.90	22.	0.61	1.078
G	1.30	0.32	0.65	1.058
H	0.80	16.	0.55	1.082
A—1	0.50	6.	0.63	1.058
A—2	0.95	16.	0.67	1.060
A—3	0.85	12.	0.57	1.053
S—1	0.96	22.	0.65	1.082
S—2	0.83	9.	0.56	1.064
S—3	1.34	25.	0.63	1.077
S—4	1.63	31.	0.62	1.068
S—5	0.70	10.	0.65	1.066
S—6	1.09	22.	0.65	1.082
S—7	1.18	28.	0.63	1.039
S—8	0.79	10.	0.61	1.038
New	—	11.	0.92	1.036
Wind Packed	—	11.	0.61	1.208

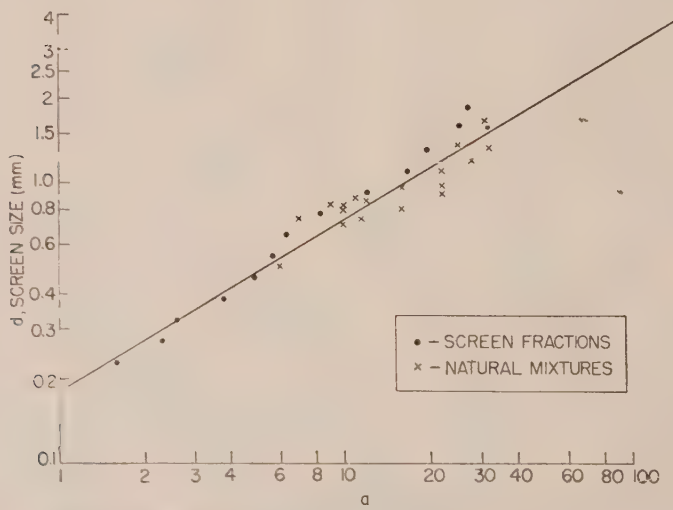

Fig. 19 — Mean grain size of snow *vs a*; Screen fractions — . ; Natural mixtures — X.

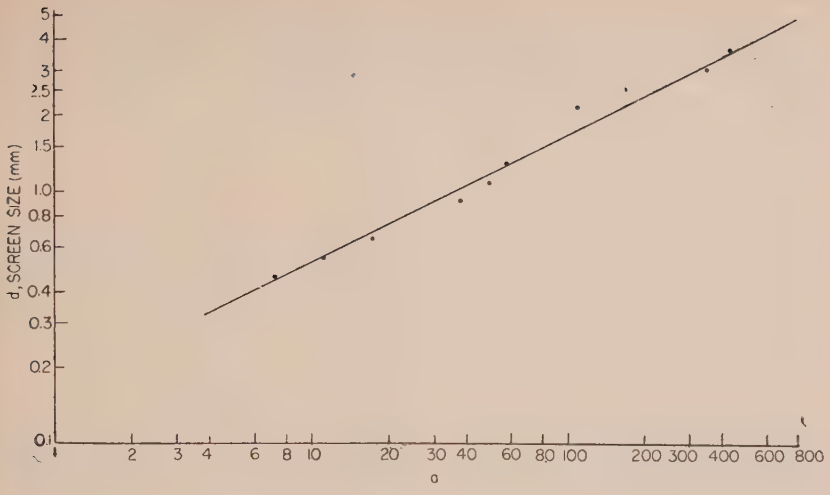

Fig. 20 — Mean grain size of glass bead fractions vs *a*.

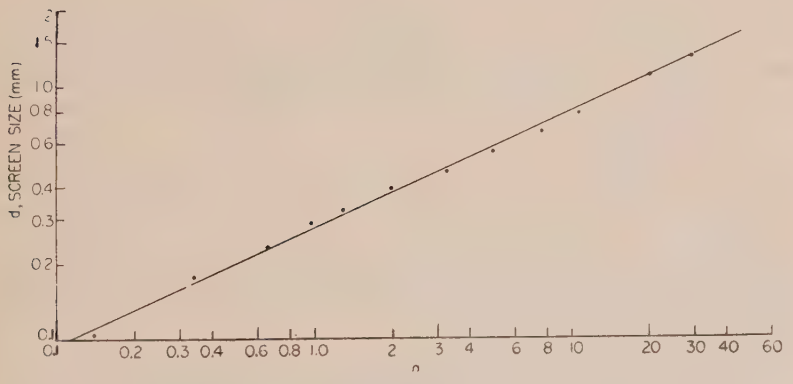

Fig. 21 — Mean grain size of soil fractions vs *a*.

It was to be expected that K for glass beads would be proportional to the square of the diameter of the beads. The values of K for glass beads are comparable to those obtained by numerous other investigators as reviewed by Carman (1956).

The meaning of N is not clear. It seems to be representative of the loosest type of packing available for the given grain shape and surface roughness of the material. For instance, the average N value is about 0.50 for glass beads; about 0.63 for the more irregular soil; and varies from as high as 0.92 for new snow to less than 0.50 for high-density old snow.

Bader (1939) noted that the ratio of N/n_0 was very nearly constant for almost all snow types and the mean value for 50 determinations was 1.058, except for extreme formations of wind crust, which had values larger than 1.1. The present investigations

show an average N/n_0 value of 1.063 and for a wind crust, 1.208. Permeability studies on snow over 10 years old (over 30 feet deep on the Greenland Ice Cap) and with densities of over 0.5 shows that this ratio still holds. The average ratio of 11 different high-density snows was 1.060. All these results give credence to the concept that N represents the loosest packing of the material for the given shape of the particles.

Conclusions

Studies on snow indicate that air flow is laminar for velocities less than 5 cm/sec for fine-grained snow (less than 0.8 mm diameter); 2 cm/sec for medium-grained snow (0.8 — 1.2 mm diameter); and 1 cm/sec for larger-grained snows. All of the results obtained under laminar flow conditions may be represented by a single relationship

$$K = \frac{16.8\, d^{1.62} n N}{N - n}$$

where K is the air permeability, n is the porosity, d is the average snow grain size in mm and N represents the loosest packing (highest porosity) for the given shape of the snow particles.

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A SURVEY OF ARCTIC SNOW-COVER PROPERTIES AS RELATED TO CLIMATIC CONDITIONS

MICHAEL A. BILELLO
Climatic and Environmental Research Branch
Snow Ice and Permafrost Research Establishment
U.S. Army

Summary

An analysis of snow cover density, temperature and hardness data measured over a period of several years at five stations in Alaska and six stations in the Canadian arctic, shows the snow cover in the Canadian Archipelago to be colder, denser and harder than in the interior of Alaska.

A series of nomographs were developed to estimate average monthly snow cover densities (g/cm^3) from mean monthly air temperature and wind velocity. The nomographs are applicable for the months November through March, for the Alaskan and Canadian area north of 62^0 N latitude, and for elevations below 1500 feet. A comparison of observed snow cover densities, with those derived from the nomographs, indicates that the method will provide a reliable regional estimate of snow cover density.

Studies of the association between snow cover temperatures and air temperature from November through March disclosed the snow to be on the average from 4^0 to 9^0 C warmer than the air at the Alaskan stations and 4.5^0 C warmer at the Canadian Archipelago stations.

An investigation of snow cover hardness revealed regional variation similar to that for density. Measured snow hardness, during the period of no melting, was found to be related to densities between 0.15 and 0.36 g/cm^3.

Preface

The data used in this report were made available through the combined efforts of personnel associated with the Canadian Department of Transport, the U. S. Weather Bureau and the U. S. Air Weather Service. Mr. Michael A. Bilello, of the Climatic and Environmental Research Branch, analysed the data and prepared the preliminary report. Mrs. Genevieve Jones assisted with the computation and tabulation of the data. The work was conducted under the general supervision of Mr. Marvin Diamond. Assistant Chief, and Dr. R. W. Gerdel, Chief, of the Climatic and Environmental Research Branch. This paper constitutes a progress report on Project 22.5-1, Snow Cover Studies.

Introduction

The purpose of this study was an investigation of the regional variations of snow cover properties in the arctic, and the development of methods for forecasting these properties from meteorological conditione.

This report covers one phase of a SIPRE research program in which the relationship between the characteristics of a snow cover and the meteorological conditions to which it has been subjected are being investigated. The objective is to provide information which can be used as a basis for increasing the efficiency of military operations or civilian activities in an environment dominated by a snow cover and low temperatures for much of the year.

A study of this type requires concurrent measurements of both snow cover properties and meteorological elements. To obtain such measurements a systematic snow observation program was initiated at various locations in Alaska and in the Canadian Arctic Archipelago, (recently given the official name, Queen Elizabeth Islands), during the winter season of 1952-1953. The observations were made at weather stations in Alaska operated by the U. S. Air Force and at the Joint Arctic Weather Stations in the Canadian Archipelago operated cooperatively by the U. S. Weather Bureau and the Meteorological Division of the Canadian Department of Transport.

Two to four years of snow cover data have been compiled for five stations in Alaska, five stations in the Canadian Archipelago and for two locations at Frobisher on Baffin Island, Northwest Territories of Canada (Figure 1).

Fig. 1
Location and elevation of stations covered in the study.

The observations which were made in accordance with instructions in SIPRE Instruction Manual ##1 (Snow Ice and Permafrost Research Establishment, 1954) provide data on certain properties of each layer of snow in a vertical profile. The different layers of snow can be visually delineated by structural differences, such as size and shape of the snow grains, or textural variations identifying periods of major snow accumulation. The properties which were measured weekly are density, temperature, hardness and crystal size.

Some preliminary studies on data from these stations indicated that there were important regional differences in the physical properties of the snow cover. In this report an effort has been made to associate these regional differences in the properties of the midwinter arctic snow cover with specific climatic conditions.

As shown in Figure 2, arctic snow cover density varies both regionally and seasonally. It is a snow property that appears to be affected by the meteorological environment.

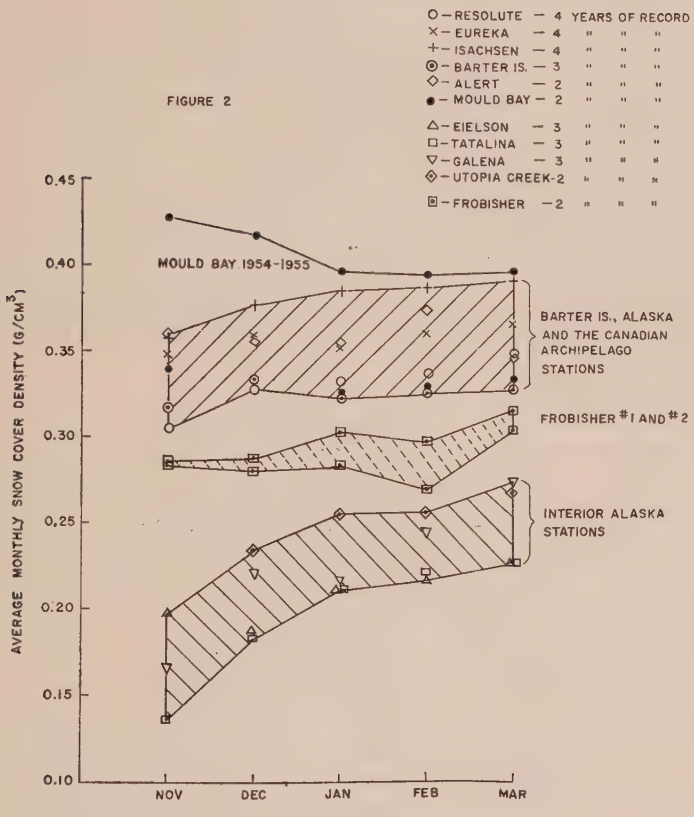


FIGURE 2
O – RESOLUTE – 4 YEARS OF RECORD
× – EUREKA – 4 " " "
+ – ISACHSEN – 4 " " "
⊙ – BARTER IS. – 3 " " "
◇ – ALERT – 2 " " "
● – MOULD BAY – 2 " " "
△ – EIELSON – 3 " " "
□ – TATALINA – 3 " " "
▽ – GALENA – 3 " " "
◈ – UTOPIA CREEK-2 " " "
⊡ – FROBISHER – 2 " " "
AVERAGE MONTHLY SNOW COVER DENSITY (G/CM³)
0.45
0.40
0.35
0.30
0.25
0.20
0.15
0.10
NOV
DEC
JAN
FEB
MAR
MOULD BAY 1954-1955
BARTER IS., ALASKA AND THE CANADIAN ARCHIPELAGO STATIONS
FROBISHER #1 AND #2
INTERIOR ALASKA STATIONS

Fig. 2
Regional and seasonal variations in snow cover density.

Those meteorological elements which may affect the density of arctic snow cover and are regularly measured at weather stations are air temperature, wind velocity, relative humidity and cloud cover.

The objective of this particular phase of the study was to determine the effect of the above climatic parameters on the density of winter snow. Since melting results in rapid and large changes in density, the study was confined to the non-melt period in the area north of 62° N latitude. This period generally extends from November through March.

TABLE 1

Mean Monthly Cloud Coverage (1.0 = *overcast*; 0.0 = *clear*)

Station	Period of Record	Nov.	Dec.	Jan.	Feb.	Mar.	Mean
Alaskan Stations							
Galena	46 — 53	.7	.7	.5	.5	.6	.6
Eielson	53 — 55	.6	.6	.6	.5	.5	.6
Tatalina	52 — 55	.8	.7	.5	.6	.6	.6
Utopia Creek	51 — 55	.7	.6	.5	.6	.6	.6
Barter Island	47 — 53	.8	.6	.5	.5	.6	.6
							.6
Canadian Archipelago Stations							
Alert	50 — 53	.5	.4	.4	.4	.5	.4
Eureka	47 — 53	.4	.3	.3	.4	.4	.4
Isachsen	48 — 53	.5	.4	.4	.5	.5	.5
Mould Bay	48 — 53	.5	.4	.4	.4	.4	.4
Resolute	47 — 53	.6	.4	.3	.4	.4	.4
							.4

TABLE 2

Mean Monthly Snow Cover Temperatures °C

Station	Period of Record	Nov.	Dec.	Jan.	Feb.	Mar.	Mean
Alaskan Stations							
Galena	53 — 56	—8.4	—15.7	—14.4	—13.2	— 9.5	—12.2
Eielson	52 — 55	—8.3	—14.4	—16.5	—11.7	— 6.2	—11.4
Tatelina	53 — 56	—7.1	—11.7	—10.3	—13.8	— 6.1	— 9.8
Utopia Creek	53 — 55	Msg.	—15.6	—13.2	—16.6	—12.2	—14.4
Barter Island	52 — 55	—9.0	—17.2	—20.8	—22.7	—21.2	—18.2
							—13.2
Canadian Archipelago Stations							
Alert	54 — 56	—22.6	—28.5	—31.4	—34.6	—33.8	—30.2
Eureka	52 — 56	—26.8	—29.6	—32.6	—32.5	—32.7	—30.8
Isachsen	52 — 56	—23.3	—26.6	—30.0	—32.8	—33.6	—29.3
Mould Bay	53 — 56	—22.3	—27.6	—28.2	— 31.7	—31.4	—28.2
Resolute	52 — 56	—19.0	—23.2	—27.1	—31.1	—27.4	—25.6
							—28.8

The average amount of cloudiness during the winter season, as shown in Table 1, indicates that the Alaskan stations have almost 50% more cloudiness than the Canadian stations. The increase in cloud cover reduces the heat loss due to long wave radiation and accounts in part for the higher temperatures of the snow cover at the Alaskan stations (Table 2).

It is generally recognized that the values of relative humidity at temperatures below 0°C as obtained with the sling psychrometer may be in considerable error. During periods when air temperatures were below minus 35°F, relative humidity was not measured at the Canadian arctic stations. For these reasons it was not possible to undertake a study of the effect of relative humidity on snow cover density.

Additional processes not directly attributable to current meteorological phenomena could effect the properties of a snow cover. The temperature gradient and thermal conductivity of the soil for example, determines the heat flow from soil to the snow and may contribute to the metamorphic processes within the snow cover. Soil conditions, such as water content and depth of freezing, which may have an influence on the snow cover, may be related to meteorological phenomena which occurred prior to the development of the seasonal snow cover. Except for some near surface soil temperatures measured at Resolute, no observations were made on soil conditions at the arctic stations. Therefore, the possible influence of soil conditions on the metamorphic processes in the snow could not be included in this report.

Air temperature and wind velocity were measured with sufficient accuracy and frequency to permit the evaluation of their effect on the density of the arctic snow cover.

Results of Study

The Regional Variations in Snow Cover Density

A weighted mean of snow cover density for each weekly observation was derived in the following manner:

	Thickness of each layer (cm)	% of total depth	observed snow density g/cm³	weighted snow density g/cm³
layer 1 (bottom)	6	20	.205	.041
layer 2	15	50	.290	.145
layer 3 (top)	9	30	.230	.069
total depth =	30 cm	100%		.255 g/cm³ weighted mean

This method of computation was used to allow for variations in thickness of the layers forming the snow cover. The density samples were taken at the midpoint of each layer, and were assumed to be representative for that layer.

Mean monthly densities were obtained from thea rithmetic means of the weekly values. These mean monthly values of snow cover density and the location, elevation and number of years of record for each station are shown in Table 3.

A plot of these values, Figure 2, reveals a marked difference in average monthly snow cover density for the interior Alaskan stations, (Eielson, Tatalina, Galena and Utopia Creek) when compared with the Canadian Archipelago stations, (Resolute,

TABLE 3

Mean Monthly Values of Snow Cover Density (g/cm³)

Station Name	Location and Elevation	Years of Record	Nov.	Dec.	Jan.	Feb.	Mar.
Tatalina, Alaska	62° 54′ N 155° 59′ W 939 Ft.	3	.136	.186	.212	.223	.229
Galena, Alaska	64° 43′ N 156° 54′ W 120 Ft.	3	.167	.221	.217	.246	.275
Eielson, AFB Alaska	64° 39′ N 147° 04′ W 547 Ft.	3	.198	.187	.212	.218	.229
Utopia Creek, Alaska	66° 03′ N 153° 45′ W 1378 Ft.	2	Msg.	.235	.257	.258	.269
Barter Island Alaska	70° 07′ N 143° 40′ W 50 Ft.	3	.319	.332	.324	.326	.329
Frobisher 1 (Weather Station Site) N.W.T.	63° 44′ N 68° 33′ W 68 Ft.	2	.284	.288	.284	.298	.317
Frobisher 2 (Runway site) N.W.T.	63° 44′ N 68° 33′ W 68 Ft.	2	.286	.282	.304	.270	.305
Resolute, N.W.T.	74° 41′ N 94° 54′ W 56 Ft.	4	.306	.330	.333	.339	.350
Mould Bay N.W.T.	76° 16′ N 119° 28′ W 50 Ft.	3*	.378	.367	.348	.350	.354
Isachsen, N.W.T.	78° 47′ N 103 °32′ W 83 Ft.	4	.359	.376	.386	.388	.393
Eureka, N.W.T.	79° 59′ N 85° 57′ W 8 Ft.	4	.348	.359	.352	.361	.367
Alert, N.W.T.	82° 30′ N 62° 20′ W 205 Ft.	2	.359	.358	.355	.375	.348

* Including 1954-55.

Eureka, Isachsen, Alert and Mould Bay) and Barter Island on the Alaskan north coast. The stations are grouped to emphasize that the study is of regional rathes then point variations in snow cover density. The range of values for the interior Alaskan stations is approximately .13 to .20 in November and .23 to .27 g/cm³ in March. The densities at Barter Island, Alaska and the five Canadian archipelago

stations range from approximately .30 to .36 in November, and .33 to .39 g/cm³ in March. Snow cover densities measured at the two sites at Frobisher, Northwest Territories range from approximately .27 to .32 g/cm³. Unusually high densities reported from Mould Bay during 1954-1955 were treated and shown separately in Fig. 2. These values range from approximately .43 in November to .395 in March. As will be shown later these observations coincided with high values in snow hardness. No extreme values of air temperature or wind velocity appeared to be associated with the se high densities.

The Prediction of Average Monthly Snow Cover Density in the Arctic from Air Temperature and Wind Velocity

Mean daily air temperatures at each station were computed from the daily maximum and minimum temperatures.

Average monthly air temperatures for these stations, computed from the mean daily temperatures are presented in Table 4. The period of record covered here is the same as that shown for the average monthly snow cover density (Table 3).

TABLE 4

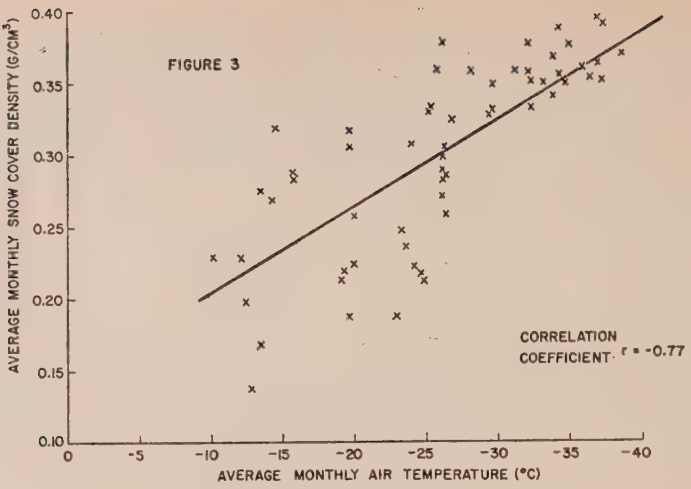
Average Monthly Air Temperature (°C)

Station	Nov.	Dec.	Jan.	Feb.	Mar.
Galena, Alaska	—13.5	—24.2	—24.6	—23.3	—13.5
Eielson, Alaska	—12.5	—23.0	—24.8	—19.4	—12.2
Tatalina, Alaska	—12.8	—19.8	—19.1	—20. 0	—10.2
Utopia Creek, Alaska	—12.5	—23.6	—20.0	—26.4	—14.4
Barter Island, Alaska	—14.5	—25.4	—26.8	—29.4	—25.2
Frobisher, Canada	—15.8	—26.1	—26.4	—26.2	—19.7
Alert, Canada	—25.9	—31.2	—32.0	—34.9	—34.6
Eureka, Canada	—29.6	—35.8	—36.4	—36.9	—38.5
Isachsen, Canada	—28.2	—32.1	—34.2	—37.3	—36.8
Mould Bay, Canada	—26.3	—33.9	—33.2	—37.1	—34.1
Resolute, Canada	—24.0	—29.6	—32.4	—33.8	—32.3

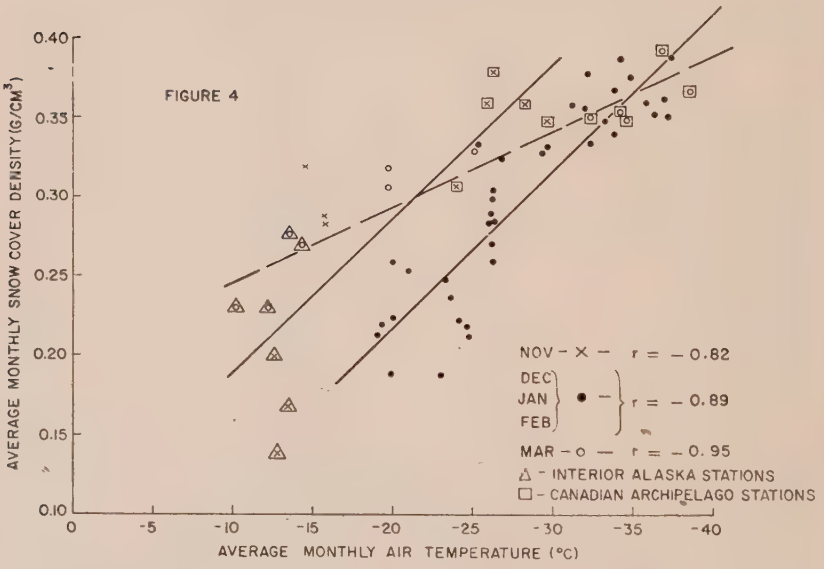
A plot of the average monthly snowc over density versus air temperature for all the stations for the period November through March is shown in Figure 3. The linear relationship based on the least squares computation yields a correlation coefficient of —0.77 with a standard error of estimate of .04 g/cm³.

A more detailed examination of the data, on a month to month basis, revealed that there were variations in the positions of the lines of best fit for certain periods of the snow season. One line fitted the data for November, another for March and a third for December, January and February. The variation is shown in Figure 4.

The November line incorporates the low snow densities and relatively high temperatures of the interior Alaskan stations and the high snow density and colder air temperature of the Canadian archipelago stations. At lower latitudes the snow accumulation season has just started by November. The newly fallen snow, not having been subjected to severe temperature changes or wind action, forms a soft snow cover of low density. In contrast, at the Canadian archipelago stations, the snow season is more advanced by November and the snow cover has been subject to meteorological influences for a longer period of time.



Relation between air temperature and snow cover density.



Relation of air temperature to snow cover density on monthly basis.

The slopes and intercepts of the regression lines for December, January and February were practically the same, indicating a similarity of response of the snow cover to climatic factors at all stations during that period. The three months, therefore, were combined into one group.

During March, the total daily direct solar radiation is greater at lower latitudes, so at the interior Alaska stations and at Frobisher on thes outhern end of Baffin Island the upper layers of the snow cover undergo a net gain in radiational heat. The absorbed radiation increases the rate of metamorphism, which results in settling, and increases the density of the pack. Oda and Kudo (1941) observed similar changes

in the snow cover, as solar radiation increased, in their studies in Japan. The process leading to increased density in the arctic snow may be similar to that described by hydrologists as «ripening» of a temperate zone snow pack prior to discharge of melt water. Since the metamorphic process has not reached this stage in the higher latitudes in March, the regression line for snow density in the more northern stations remains relatively unchanged.

For the area covered by this study, the following relationships between snow density and air temperatures appear to apply.

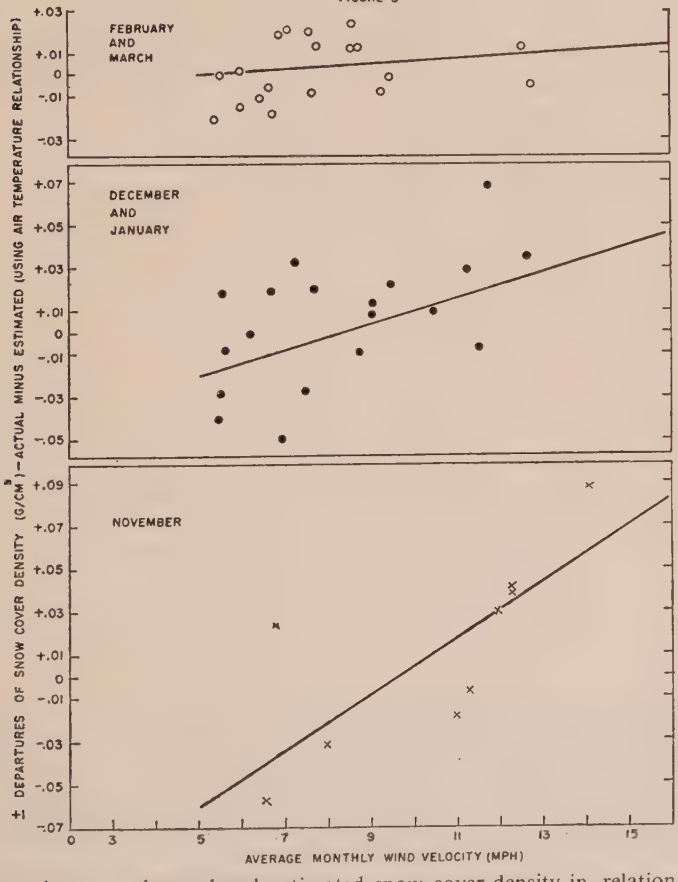
For November: $\rho_s = -.0098T_A + .090$

For (December:) (January:) (February:) $\rho_s = -.010T_A + .016$

For March $\rho_s = -.0048T_A + .197$

where: ρ_s = average monthly snow cover density (g/cm³).
and: T_A = average monthly air temperature (°C).

FIGURE 5


FEBRUARY AND MARCH
DECEMBER AND JANUARY
NOVEMBER
± DEPARTURES OF SNOW COVER DENSITY (G/CM³) — ACTUAL MINUS ESTIMATED (USING AIR TEMPERATURE RELATIONSHIP)
AVERAGE MONTHLY WIND VELOCITY (MPH)

Difference between observed and estimated snow cover density in relation to wind velocity. (Densities estimated using air temperature).

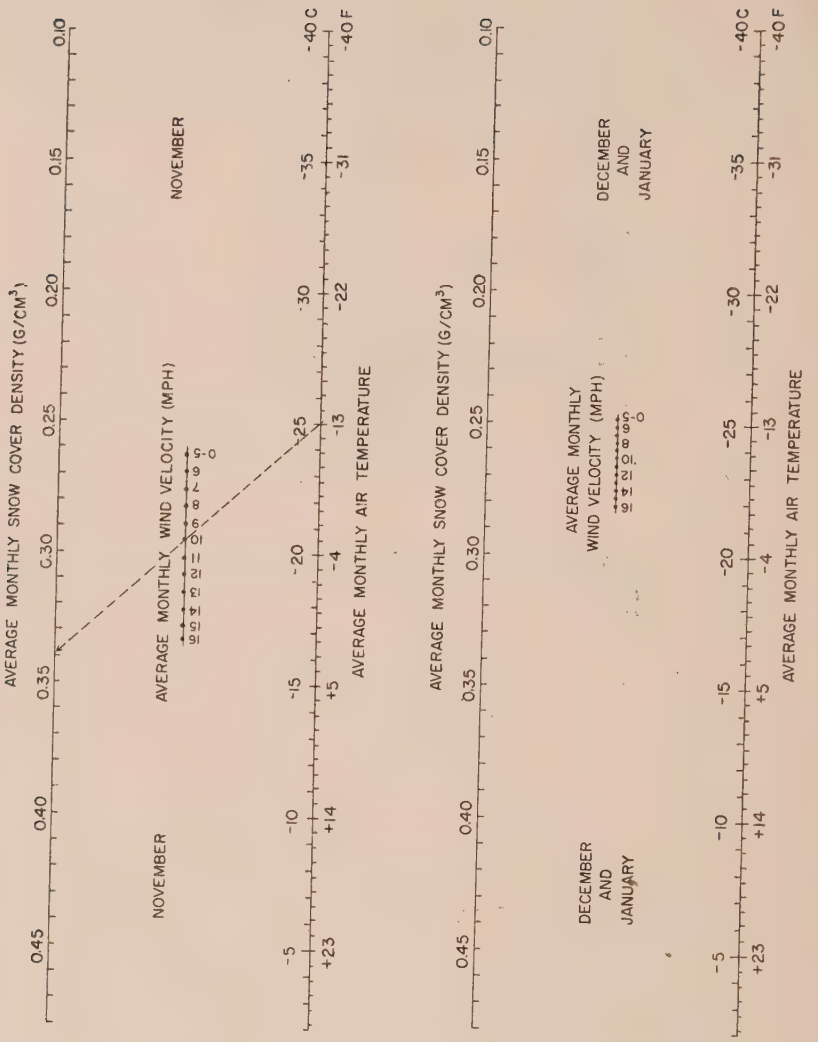
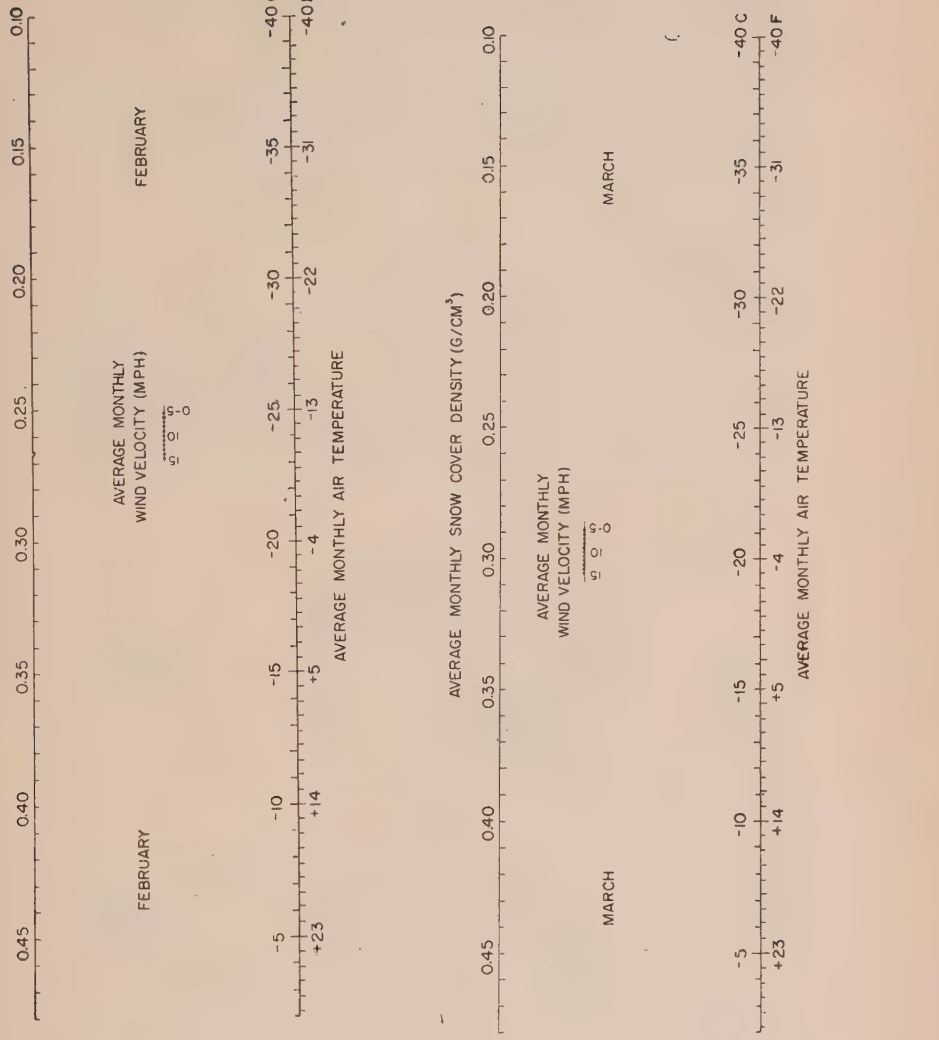


Fig. 6A through 6D — Nomographs to estimate average monthly snow cover density for November through March.

Exemple : Assume average monthly air temperature for November is -25 °C and average wind velocity is 10 MPN. cln. Fig. 6A, extend a straight line through these 2 points and read 342/cm³ on the average monthly snow cover density scale.



An investigation of possible causes for departures of the actual snow cover density from values estimated from the regression equations, indicated that high snow densities frequently were associated with strong winds and low densities with light winds.

A comparison of these departures from estimated values with the corresponding average monthly wind velocity is shown in Figure 5. Wind velocity appears to be more influential in November than in February and March. This seasonal decrease in the apparent effect of wind velocity on snow cover density is associated with, 1)

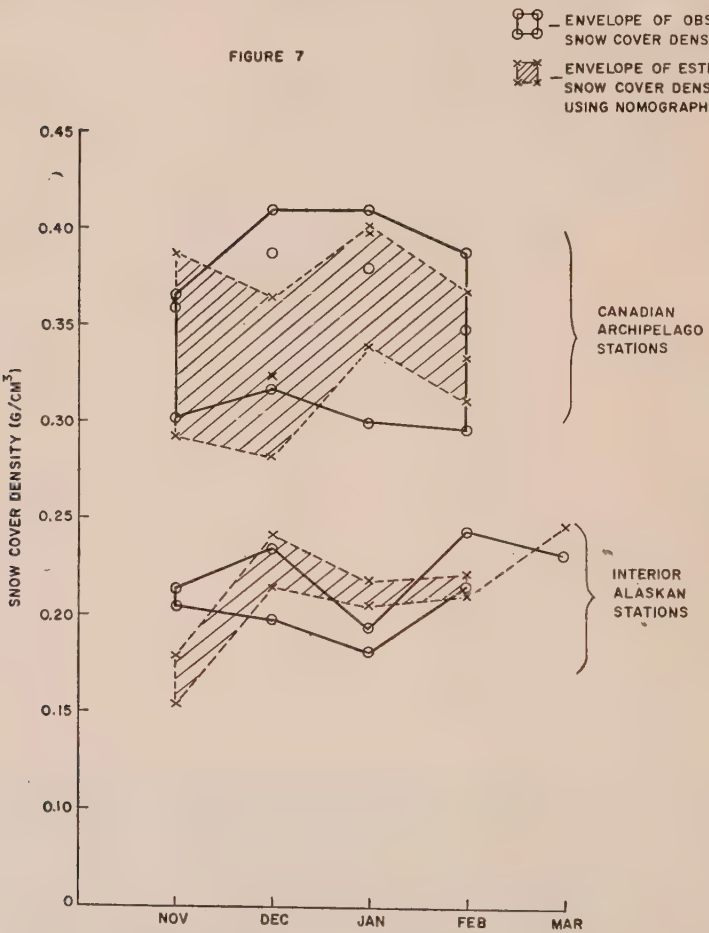
the period when most of the snow has accumulated and, 2) with the seasonal increase in snow cover density. Studies show that in the area under discussion, 52% of the mid-winter snow is deposited by November 30, 80% by January 31; and the snow cover density increases from an average of .28 by the end of November to .32 g/cm³ by the end of March.

Since both air temperatures and wind velocity affect the snow density, these variables were used to develop a method for graphically computing average monthly snow cover densities. These nomographs (Figures 6A through 6D) for November through March are applicable to the Canadian and Alaskan regions north of 62° N latitude, and for elevations below 1,500 feet.

Comparison of Observed Snow Cover Density with Those Computed from the Nomographs

Meteorological data for Resolute, Eureka and Alert in the Canadian archipelago during the 1956-57 season has been used to compute densities from the

FIGURE 7



Comparison of observed snow cover densities with densities computed from the nomographs, for 1956-1957.

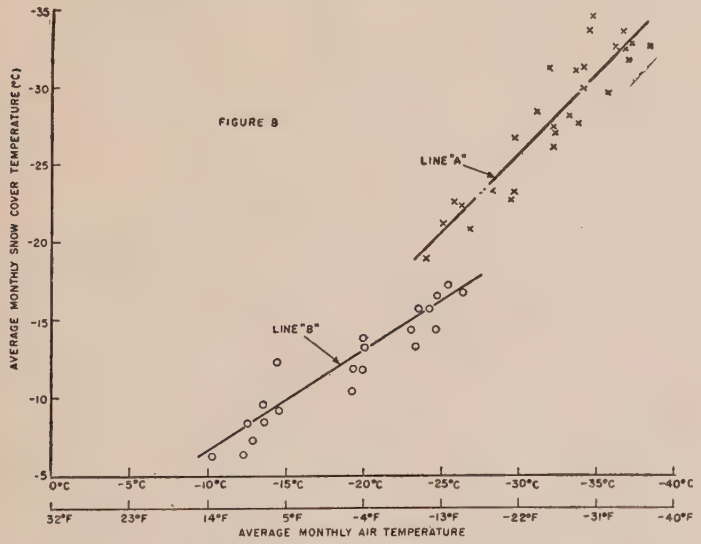
nomographs. The estimated densities and the observed densities (values which were not used previously) for identical periods are shown in Figure 7.

On the same chart observed densities at Tatalina and Utopia Creek, Alaska also, for the 1956-57 snow season, are compared with computed densities from climatological data obtained from the weather records published in the Air Weather Service's. «Uniform Summaries of Surface Weather Observations». The climatological reports were used because the weather data for 1956-57 was not available.

There is fair agreement between the actual and estimated snow cover densities on the basis of regional variations. However, for individual locations within regions, the accuracy of prediction decreases.

The Relationship between Average Monthly Snow Cover Temperature and Air Temperature

The mean snow cover temperature for each observation was computed by weighting the temperature of each layer by the ratio of the layer thickness to the snow cover thickness. The procedure was similar to that described earlier for computing the mean density of the snow cover. The mean monthly values of snow cover temperature computed arithmetically from the weekly means are listed in Table 2. The relationship between monthly snow cover temperature and the corresponding monthly air temperatures (Table 4) is presented graphically, in Figure 8.



Relation between air temperature and snow cover temperature

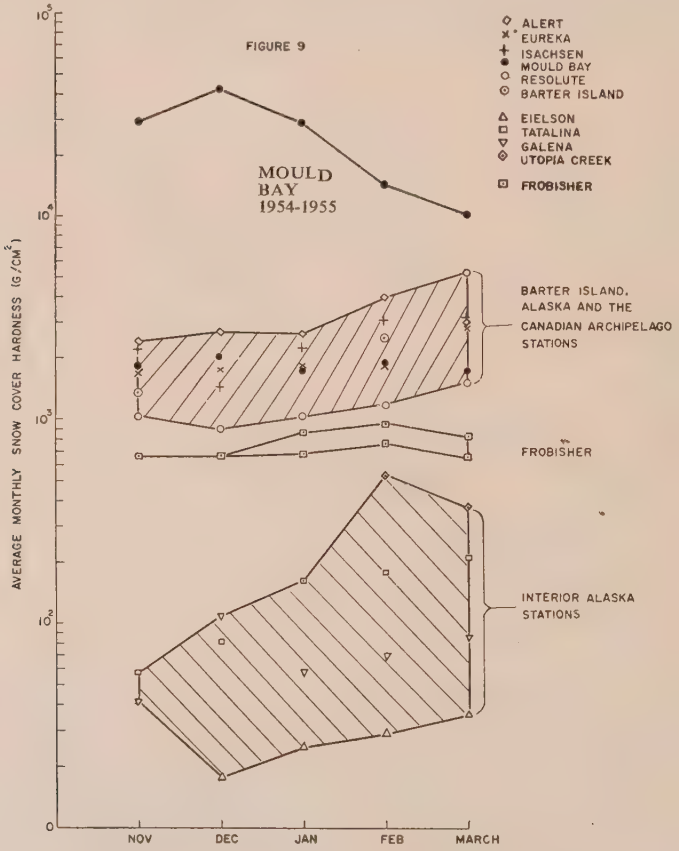
The distribution of points in this figure indicates that the data used in this study probably were derived from two separate populations. Line A appears to define a climatic regime in which the snow temperatures are about 4.5°C warmer than the air temperature while line B defines a regime where snow temperatures are from 4°C to 9°C warmer than the air temperature. The data to which line A has been fitted were all derived from stations located north of 70° N latitude, and with but two exceptions (Barter Island, Nov. and Dec.) all points assigned to line B were derived from records for stations located between 63° and 66° N latitude.

The difference between the two climatic regimes as indicated by these lines may be due in part to differences in the physical properties of the snow cover, in part to differences in soil properties and heat storage in the soil, and as previously discussed, the amount of cloudiness. The possible influence of soil temperature on the snow cover temperature is indicated by the —0.2 to —0.4 °C geotherm found from 30 to 60 ft. depth at Ladd, Alaska, near latitude 65°; whereas at Point Barrow, Alaska, latitude 70° N and Thule, Greenland, latitude 76.5° N a soil temperature of —9.5°C is found from 30 to 200 ft, below the surface. Sufficient data on soil temperatures are not available for the stations included in this report to determine differences in magnitude of heat flow between the earth and the snow cover. It appears, however, that differences in snow cover temperaturesm ay be due in part to differences in soil temperatures.

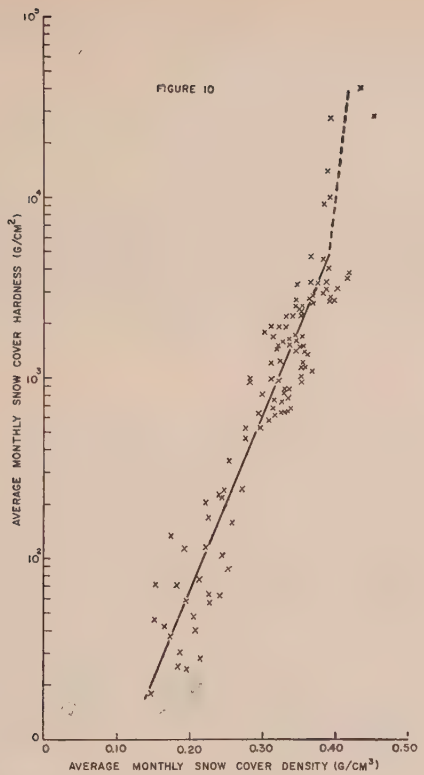
The Relationship between Hardness and Density of the Snow Cover

The geometric mean was used to compute mean snow cover hardness in order to reduce the effect of extreme values.

The data plotted in Figure 9 revealed a regional and seasonal variation in the mean monthly hardness of the snow cover similar to that of the density



Saesonal and regional variations in snow cover hardness



Relation between density and hardness of the snow cover.

(Figure 2). This indicates an association between the hardness and density of the snow cover. The relationship is shown in Figure 10.

The line of best fit extends without change in slope to a hardness of 5000 gm/cm^2 The plotted points above that value are for the hard and dense snow cover reported from Mould Bay during 1954-1955.

Increases in the measured hardness of the arctic snow cover may be attributed to two processes. One involves an increase in the number of snow grains per unit volume, which is accompanied by an increase in density, as indicated in Figure 10, for densities between .15 and .36. The second is due to the development of ice bonds between the snow grains and would not be accompanied by an increase in snow density. The instruments and procedures used in measuring hardness do not permit separate identification of the two processes. Consequently, it was not possible to establish relationships between snow hardness and the meteorological parameters identified with snow cover density.

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SNOW SURVEYING BY ONTARIO HYDRO

J. A. S. MILNE

Abstract

Snow surveys have been conducted in the Province of Ontario by Ontario Hydro personnel since 1931. This paper briefly describes Ontario Hydro, its formation and function; the topography of the Province of Ontario; the need for snow surveys; description of a snow course; type of equipment used; its component parts; snow survey procedure; and finally the use of survey data present and future.

General

The Hydro-Electric Power Commission was established in 1906 by an enactment of the Ontario Legislature. This provincial «Act to Provide for the Transmission of Power to Municipalities» received Royal assent on May 14, 1906. The Commission Hyphenated is a corporate entity, a selfsustaining public concern endowed under the Power Commission Act with broad powers to produce, buy and deliver electricity throughout the Province and to exercise certain regulatory functions with respect to the large group of municipal electrical utilities which it serves. This enterprise, administered by the Commission, is generally referred to as the Ontario Hydro.

Ontario Hydro operates 65 hydro-electric and two major thermal generating stations. At the end of 1956 our installed capacitya mounted to 4,552,100 kilowatts which supplies 90 per cent of the primary energy consumed in Ontario. Ten separate projects are newly authorized or under construction, with a capacity of 2,300,000 kilowatts which will bring our total installed capacity up to almost 7,000,000 kilowatts during the next four-year period.

Introduction

The Province of Ontario covers a total area of 412,600 square miles of which approximately 88 per cent is land area and the remainder lakes and rivers. The province extends 1,000 miles from east to west and 1,050 miles north to south. The mean annual precipitation varies from 25 inches in the western portion of the province to 30 to 35 inches in the eastern and southern portions. The mean annual snowfall in the western extremity of the province is 70 inches increasing easterly to 120 inches on the north shore of Lake Huron and decreasing to 40 inches in the extreme southern portion of the province.

The topography of the province is quite varied. An extensive area of land bordering on James Bay and Hudson Bay is low and swampy. A height of land runs generally east and west just north of Lake Superior; elevations in this region are in the neighbourhood of 1,000 feet and over. The land slopes very gently from this height of land to Hudson Bay and James Bay. The eastern portion of Ontario is relatively low level land draining into the Ottawa and St. Lawrence rivers. Most of this area is between 200 to 300 feet above sea level rising 400 to 500 feet north of Lake Ontario. The highest land areas in the southern portion of the province lie immediately south of Georgian Bay with elevations ranging from 1,400 to 1,700 feet. A remarkable feature is the Niagara escarpment with a height of 250 to 300 feet and extending from the Niagara peninsula northeast to the Bruce peninsula.

(As a matte of interest this escarpment in the Niagara Falls area has made possible Ontario Hydro power installations of upward of 2.4 million horsepower).

As previously mentioned the mean annual snowfall varies from 40 to 120 inches in the province. In that area of the province north of the 45° parallel of latitude the snow normally accumulates through the long winter months and then with the progress of the 32° isotherm northward melting takes place commencing March through May. The subsequent runoff, in the main, fills the storage basins which are used to generate hydro-electric energy. To the Ontario Hydro this area in question represents over 2 million installed horsepower. Because of this productive capacity it is essential that reasonable estimating of the amount of this runoff by made available before it occurs. Only by an intensive program of snow surveys, coupled with sound statistical runoff-snow relationship, can this runoff amount be determined. With this in mind snow surveys have been made continuously by Ontario Hydro personnel since 1931.

Snow Courses

Since the first surveys were initiated in 1931 the expansion of the survey has been gradual, but steady. By 1936 observations were made at 24 locations throughout the province. This past winter regular measurements were made at 57 courses throughout the province. At 35 of these courses Hydro personnel double as snow surveyors, 7 courses are surveyed by local personnel trained by Generation Department snow surveyors and 12 courses are surveyed by personnel of the Provincial Department of Lands and Forests, also trained by Generation staff.

All snow courses operated by Ontario Hydro consist of 31 observation points, 100 feet apart, along a 3,000-foot course over terrain which has been selected, by ground and air survey, as being representative of the watershed in which the surveys are to be made. Each observation point is marked with aluminum tags nailed loosely, in order that tree growth will not explode the tag from the tree, to the tree nearest each observation point. The course need not be a tangent or straight line and is chosen in an area which will remain unchanged for years to come.

Equipment

The type of equipment presently in use is adapted from the Mount Rose snow sampler. Where the Mount Rose is steel throughout, the type we use (sometimes referred to as the Federal equipment) is made of duraluminum tubing. It comes in 2-1/2-foot sections, into the lower end of the first section of which is shrunk fit a 1.485-inch (inside diameter) case hardened, cold rolled steel snow cutter. Tube sections may be assembled in various lengths as each section is made with threaded couplings of duraluminum. In Ontario our maximum depth of snow has yet to exceed 5 feet, consequently our snow survey sets are only provided with two lengths of tube. Each tube has a number of offset slots throughout its length which provides a means for examining each core for length and snow structure. The outside of the tube is marked off in inches and half inches to provide depth readings.

The Balance

The balance which is also made of duraluminum is tubular in shape and short for convenience in carrying and handling. It requires a minimum length of 5 feet

of tubing to register on the scale and has a weighing capacity of 12.5 feet of tubing. The balance is graduated to weigh in ounces and since the snow cutter cuts a snow core 1.485 inches in diameter, the scale reading is equivalent to depth in inches of the water in the sampled snow cover.

Wire Cradle

A wire cradle is provided for supporting the tube while it is being weighed.

Wrenches

A pair of spanner wrenches are also furnished with the equipment. They are steel and are used for uncoupling the sampler. Holes into which they fit are required in both ends of each coupling.

Driving wrenches are required in great depths of snow and as such depths are not encountered in Ontario these are dispensed with.

Canvas Carrying Cases

Canvas cases of fireproof and rot-proof canvas are provided with individual pockets for each part of the equipment. Straps are attached to the case which permits carrying similar to a golf bag. This leaves both hands free.

Care of Equipment

Snow sampling equipment is designed to withstand rough traetment. Neverthless, in some respects it is delicate and should have the best care possible. As well as care in the field, once a year the equipment is returned to head office where the scales are checked for accuracy and the whole equipment inspected, repaired where necessary, and returned to the field station prior to the next year's survey.

Snow Survey Procedure

Briefly the procedure for taking a snow survey is as follows:

At each of the 31 observation points in a snow course, the sampler tube is carefully pushed, perpendicularly into the snow mantle, with a slight twisting action to cut through the crusts. When the ground level is reached the snow depth is observed, using the graduations on the tube, and recorded. The tube is pushed a little further into the ground in order to pick up an earth plug. This ensures that the snow core remains in the tube when the tube is being removed from the snow pack. This is particularly important late in the winter when the snow is granular in structure. After the tube is removed from the snow the earth plug is removed and the weight of the tube and snow core in ounces is determined. The weight of the empty tube, measured before obtaining a core, is subtracted from the weight of tube plus snow thus yielding the weight of the snow in ounces and water content in inches. Before weighing the core it is examined to determine its general characteristics. On the average the core length should be at least 75 per cent of the snow depth. Compaction of the snow in the tube causes this difference.

After completing the 31 observation points averages are taken of depth and water content and these figures represent the depth and water content for the course.

Field note paper and books are provided and a recent innovation is a covering sheet which lists various data regarding weather, air temperature, etc., which the surveyor fills out before returning to his base. At home base a copy of his results are forwarded immediately to head office. This is done by mail and in isolated locations where mail service is slow, radio and telegraph is also used to transmit the results.

Times of Surveys

Snow surveys are taken once or twice a month commencing in the early winter — generally January 1st. At the end of the season, however, from the start of the snow melt period to its conclusion, surveys are taken as frequently as once a week. This is of considerable aid in regulating the rate of storage build up in the Commission reservoirs.

Use of Snow Surveys

No final equations have been derived at present for the snow water content versus spring runoff because the record for the most part is still of too short a period to yield any reliable statistical parameters. However, for the records available, averages have been computed which are used as an indication of spring supply in conjunction with other observed natural phenomena.

A further use is being made of snow survey data and that is the injection of the weighted mean water content from a given watershed into a statistical precipitation-runoff relationship. This is usually done by replacing one of the months of effective precipitation by the weighted mean water content of the first of the month following. This has proven quite successful in one of the Commission's watersheds and as each year goes by it is hoped, that with a goodly accumulation of snow data, the role of snow in the spring runoff will become more clearly defined.

Conclusions

To the hydrologist, most of his water problems arise from a lack of knowledge of future precipitation. In such a climate as ours, where the bulk of the spring runoff is from a snow pack which has been accumulating all winter, the effect of rainfall at the time of snow melt is also a major factor in producing excessive runoff conditions. The role of the meteorologist in this regard is of the utmost importance. Long-term precipitation forecasts of reasonable accuracy must be made available to the hydrologist in order to arrive at a more positive and efficient forecast. With the aid of high-speed computers it has now become a relatively simple and mechanical operation to integrate a forecast of precipitation with runoff on a statistical basis. Herein lies the challenge!

MEAN DURATION AND ACCUMULATION OF SNOW COVER IN CANADA

J. G. POTTER. (*)
Meteorological Branch, Department of Transport, Toronto.

ABSTRACT

With the numerous meteorological observing stations established in Northern Canada during the war years, and the more detailed observing program for snow cover adopted in 1941, data are not available for the first time on the mean duration and accumulation of the winter's snow cover. Data for 15 winters serve as the basis for discussion of the general features of the duration and depth of snow cover in Canada. Among the charts included are those illustrating the mean dates of the beginning and ending of the winter snow cover, annual duration of snow cover, and the mean accumulation of snow on the ground.

Routine daily measurements of the depth of snow cover was begun at first order weather observing stations in Canada in 1941. Prior to that year the amount of freshly fallen snow was measured regularly at all stations, while the accumulation of snow on the ground was only recorded each monday morning and at the end of the month by the majority of stations. Any attempt to describe snow cover over all of Canada prior to 1941 is also impossible due to the lack of data from northern regions. Many of the observing stations in the northern sections of the mainland and in the southern Arctic islands were established about this time and, with the establishment of the Joint Arctic Weather Stations in the northern islands after the end of World War II, data from all parts of Canada became available.

In measuring the depth of the snow cover the observer is instructed to take a series of measurements in an area chosen with a view to avoiding drifts and report the average depth in inches. At exposed observing sites, especially in the prairies and in the Arctic where strong winds may drift the snow into gullies and the lee of obstructions leaving most of the ground relatively bare, it is extremely difficult to make representative observations.

After daily observational data on snow cover for ten winters became available they were summarized and the main features of the results presented (1). The following description is based on data for fifteen winters beginning with that of 1941-42. In the figures the isolines are based on data from regular weather observing stations. In the mountainous areas of western Canada they will be generally representative of valley conditions, and no attempt has been made to include the rapid change with height which occur in such regions. Due to the scale of the figures it is also impossible to include the permanent snow fields which exist on the higher elevations of the Arctic Islands and in the western mountains.

In Fig. 1, isolines are drawn for the beginning of the season of snow cover. This first accumulation of one inch or more of snow on the ground may melt away and form anew once or several times before the continuous snow cover of mid-winter appears. The mean date of occurrence of the first snow cover varies by four months throughout Canada. In the northern half of the Arctic Archippelago September is the month with the greatest mean snowfall, although heavy snowfalls may occur

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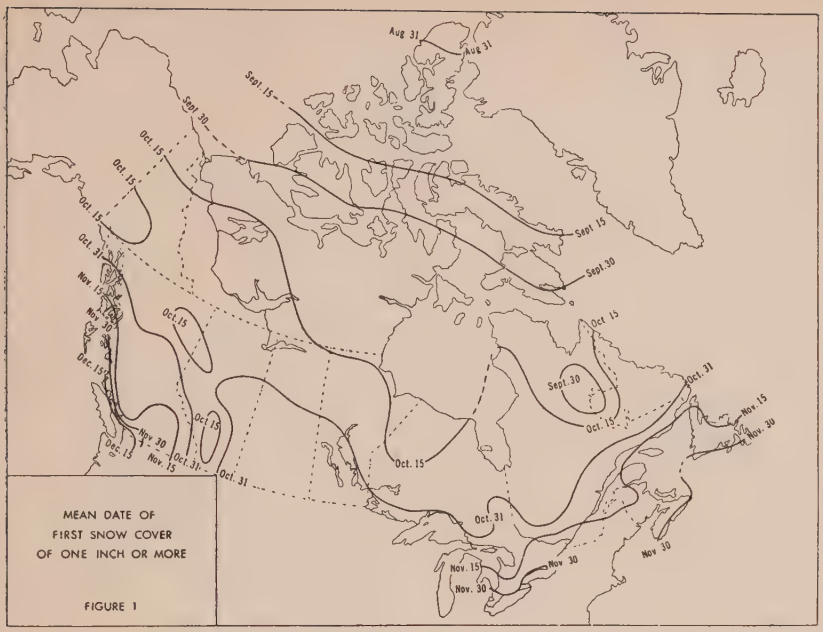

FIGURE 1

during the summer months of July and August. The mean date on which the snow on the ground first accumulates to a depth of one inch at Alert, on Ellesmere Island, the most northerly weather station in Canada, is August 25th. By the end of September almost all the Arctic Islands are snow covered.

During October the return to winter conditions, as indicated by the first snow cover, occurs in over one half of Canada. Between Hudson Bay and the Rockies the time of first occurrence is remarkably uniform. The difference in the time occurrence of the first snow cover along the Arctic coastline and the southern border of the Prairie provinces — almost 20^0 of latitude — is approximately five weeks. In the foothill country of Alberta the higher elevations experience early heavy snowstorms which give a temporary snow cover. Thus at Calgary, in the foothills in southern Alberta, the first snow cover occurs just twelve days later than at the mouth of the Mackenzie River within the Arctic circle, over one week earlier than in the vicinity of Great Bear Lake, and slightly more than three weeks earlier than over the prairies at the same latitude.

In eastern Canada, owing to differences in elevations and the warming effect of large bodies of water, the pattern is more complex. By the end of September the first snow cover is likely to have occurred on the northern slopes of the interior highlands of Quebec. At the same latitude along the shore of Hudson Bay the relative warmth of the waters in Autumn retards the formation of snow cover until two weeks later. In the more densely settled areas of eastern Canada the first snow cover is normally observed in November, except on the shores of the lower Great Lakes and along the Atlantic coast of Nova Scotia, where it does not normally occur until early in December.

The mean accumulation of snow on the ground near mid-winter is illustrated in Fig. 2. The most striking feature is that there are very few sections of Canada which do not have more snow on the ground in mid-winter than the islands of the

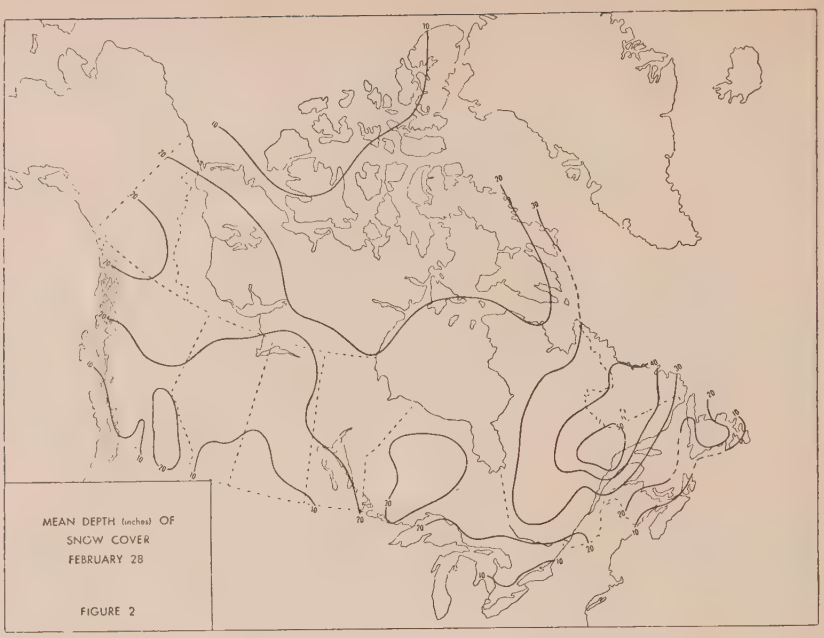

FIGURE 2

Arctic Archipelago bordering on the polar sea where the normal accumulation by the end of February is less than 10 inches. Over the barren ground west of Hudson Bay and the three Prairie provinces the snow cover is remarkably uniform. It is normally close to twenty inches in depth. From Ontario eastward the winter precipitation is greater due to the more frequent passage of cyclones with warm moist air in their warm sector. While precipitation is greatest in the southeast over the maritime provinces and St. Lawrence valley, the accumulation of snow on the ground in these sections is frequently diminshed by periods of melting. Thus the greatest accumulation of snow occurs north of the St. Lawrence valley and Great Lakes is in a region with lower winter precipitation, but with temperatures low enough to insure that the precipitation falls as snow and is not likely to be depleted by melting. It averages over 40 inches in parts of northern Ontario and over 50 inches in parts of central Quebec.

The average of the greatest depth of snow observed during each winter has been illustrated in Fig. 3. This greatest accumulation usually occurs in January along the Pacific coast, during February in the remainder of southern Canada, and later at higher latitudes, until in the Arctic Archipelago the greatest snow depth of the winter usually occurs in May.

In the Arctic Islands, except for the eastern side of Baffin Island, the mean maximum depth is between 10 and 20 inches and depths greater than 2 feet are very unlikely. Storms passing northward through Davis Strait cause heavier snowfall on the shores of Baffin Island. There the average maximum snow depths vary from 20 to 40 inches, and have exceeded 5 feet in unusual winters.

In the extensive region from Hudson Bay to the Rockies the means of the maximum depths are again fairly uniform. The greatest depths observed during fifteen winters are close to 3 feet throughout this area, except in southern Alberta and Saskatchewan where they are closer to 2 feet, and nearer 4 feet on the shores of

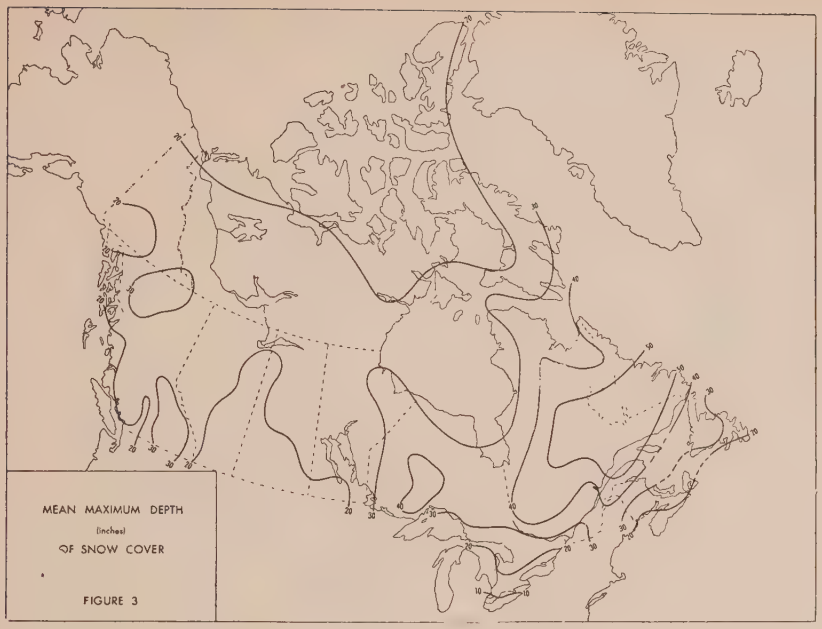

FIGURE 3

Hudson Bay and central Manitoba. The mean maximum depths are greater over most of eastern Canada where the accumulation of snow in a normal winter is greater than 40 inches northeast of Lake Superior and greater than 50 inches over large areas in northern Quebec and Labrador. Extreme depths of 5 feet have occurred in northern Ontario and from 7 to 8 feet in Quebec and Labrador north of the Gulf of St. Lawrence. Fig. 3 does not show the region of the greatest accumulation, as large areas of the higher elevations of the coast mountains in British Columbia are covered each winter by more than 30 feet of snow.

The change from winter to spring conditions with the melting of the snow does not coincide in Canada with the official change of season in March. Fig. 4 illustrates the mean date of the last snow cover of one inch or more. In most regions this date would be one to two weeks later than that of the disappearance of the continuous snow cover which exists throughout the winter over most of Canada. It is usual for the continuous snow cover of winter to be almost melted over the agricultural areas of the Prairies by the end of March, but snowstorms occurring during the spring give a mean date for the end of the snow cover season from the middle to the end of April. As in the beginning of the season, there is an unusual uniformity over many degrees of latitude in the end of the season of snow cover over the Prairies and in the Mackenzie Valley.

In eastern Canada the end of the season of snow cover is delayed much later than over the Prairies at the same latitude — the result of greater accumulation during the winter and lower spring temperatures due to more frequent outbreaks of Arctic air in spring. Its northward shift is very slow during April and the first half of May when the deep accumulation of snow over Quebec and northern Ontario is melting. With increasing hours of sunshine in late May and June the last snow cover disappears over large areas in northern Canada. By July 1st snow will normally

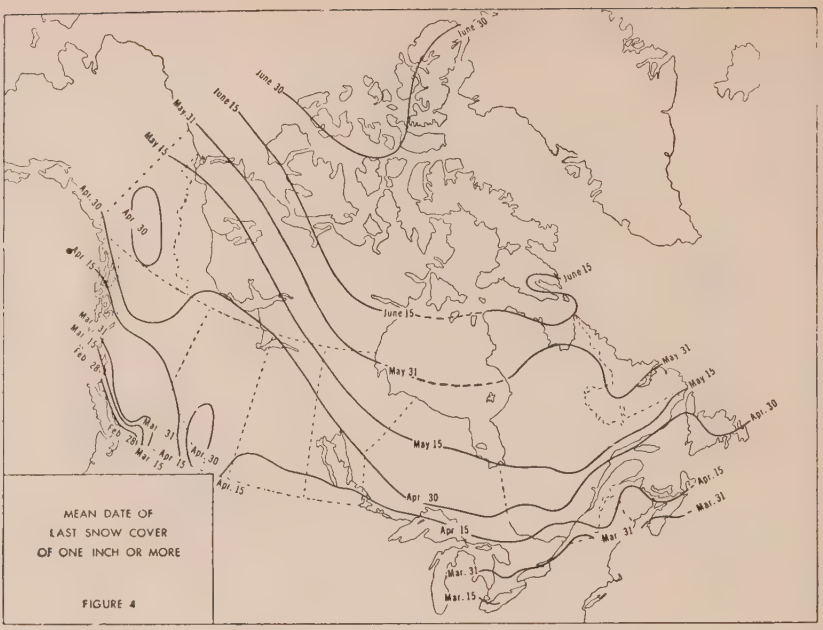

FIGURE 4

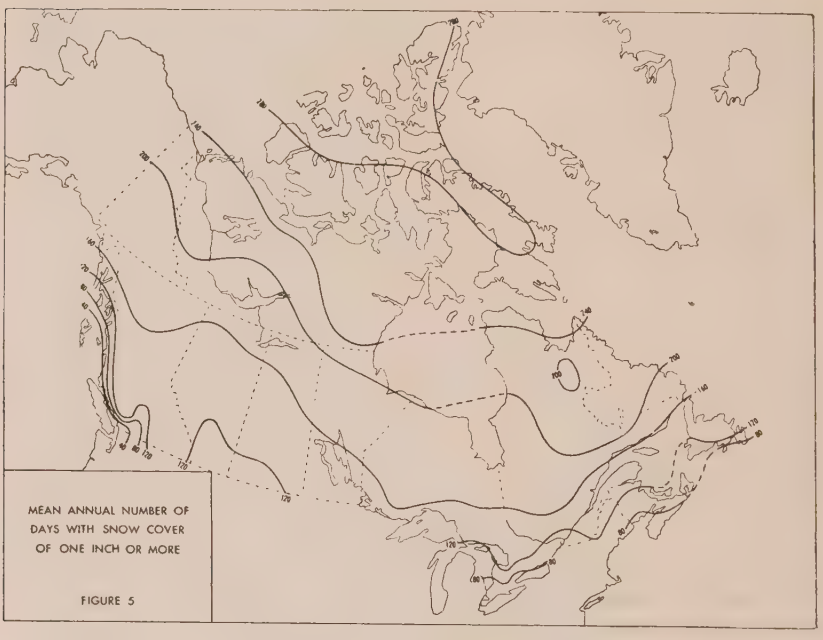

FIGURE 5

be found only on the islands adjacent to the Polar Sea and in the permanent snow fields elsewhere.

Unlike Fig. 1 and 4 which were based on the beginning and ending of the season of snow cover, Fig. 5 is based on the mean annual number of days with one inch or more of snow on the ground. There are regions along the Pacific coast where one half the winters pass without one inch of snow accumulating on the ground, while along the outer fringe of the Arctic Archipelago the ground is free of snow for about 70 days during their short summer. The duration of snow cover is greater in eastern Canada than at the corresponding latitude in western Canada just east of the Rockies. For example, the isoline for 160 days which dips southward to 50° in Quebec and Ontario is approximately 10° farther north in Alberta. If the isoline for snow cover for one half the year had been included on this figure it would have nearly divided Canada into two equal parts depending on whether the ground is snow covered or bare for the greatest portion of the year. In eastern Canada such a division extends from just north of the Gulf of St. Lawrence to James Bay, while in Western Canada almost all of the provinces of British Columbia, Alberta, and Saskatchewan are without snow cover for more than half of the year.

It was noted earlier that these figures are based on data for 15 winters, while that for the first 10 winters of the period had been similarly mapped. The addition of data for the past 5 winters has not significantly changed the results and it is hoped that the mean conditions portrayed here are thus close to normal.

In this general discussion of snow cover in Canada, no mention has been made of the water content of the accumulated snow. The regular data collected by the Meteorological Branch do not include this measurement. The National Research Council, with the co-operation of the Meteorological Branch, has carried out surveys of the physical characteristics of the snow cover and from them information on the average density is available. ([2])

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MEASUREMENT OF THE SNOW-WATER-CONTENT WITH THE USE OF RADIOCOBALT

ING. JAROSLAV MARTINEC
The Hydraulic Research Institute in Prague

ABSTRACT

A new measuring method based on the absorption of gamma-rays by water is investigated to improve the quality and quantity of data of the snow-water-content in the mountain regions, which are needed for operation of the reservoir system on the Vltava river in Czechoslovakia.

In recent years experiments in the Krkonose-mountains have been carried out. For a given arrangement of the measuring device the absorption coefficient for water has been found to be a constant for all snow-layers up to the water equivalent of 52 inches (132 cm). The values of this coefficient were similar to those obtained in California, where this method has been first adopted, and very close to the theoretical value.

In the winter season 1955-1956 the experimental snow-gaging station in the height of 4600 feet (1410 m) was in action during 5 months with the whole construction snowed up for some time.

In order to examine the accuracy of the results continuous measurements of the gamma-radiation and of the cosmic background were made and the statistical deviations of the radiation investigated. Relationships between the duration of a single measurement, the intensity of the gamma-radiation and the accuracy of the result have been derived.

INTRODUCTION

In many parts of the world predictions of the runoff from snowmelt have been used for operation of multi-purpose reservoirs. With the construction of the reservoir system on the Vltava river in Czechoslovakia in progress, forecasts of the streamflow from the water stored in the snow packs are studied. As it is often the case, only limited data of the snow-water-content in the mountain areas could be obtained by standard procedures (the melting or weighing method). Therefore a new method using the principle of absorption by water of gamma-radiation is investigated.

THE PRINCIPLE OF THE METHOD.

The principle may be briefly summarized as follows: The gamma radiation passing through a snow pack is absorbed in accordance with the equation:

$$J_x = J_o \, e^{-\mu x} \tag{1}$$

where J_o is the intensity of the incident radiation
J_x is the reduced intensity of the radiation after penetrating the snow pack
μ is the absorption coefficient for water
x is the water equivalent of the snow pack

Thus it is possible to express the measured intensity of the radiation in terms of the snow-water-content regardless to the density of the snow pack.

The theoretical approach to the problem has been outlined in the article of R.W. Gerdel, B.L. Hansen and W.C. Cassidy ([1]) in which the first application of this method in California was described. This paper contains experimental results

concerning especially the values of the absorption coefficient and the accuracy of measurements.

Results of experimental measurements.

Preliminary investigations were carried out in spring 1955 in the Krkonoše mountains in the height of 1150 m (3800 feet). The radioactive source, 15 millicuries of the radioisotope Co 60, was installed beneath a 100 cm (40 inch) snow pack. The radioisotope was enclosed in a lead shield which was so designed, that the collimated beam of the radiation penetrating the snow pack was only slightly larger than the Geiger-Müller tube, which was suspended above the snow pack in the distance of 200 cm (6 ½ feet) from the radioactive source. The intensity of the radiation was measured by counting the pulses of the Geiger-Müller tube on a decimal scaler. The snow pack between the radiaoctive source and the tube was gradually reduced and the corresponding snow layers were melted into water for comparison. Results of these measurements are presented in Table 1.

TABLE 1

Snow-water-content measurements in the Krkonoše mountains in 1955

Depth of the snow pack		Water Content measured by melting		Counting rate per minute
cm	inch	cm	inch	
100	39,4	57,7	22,6	1091
90	35,5	52,4	20,6	1574
80	31,5	46,6	18,4	1884
70	27,6	40,2	15,8	2413
60	23,6	34,8	13,7	3112
50	19,7	28,5	11,2	3951
44	17,4	25,0	9,8	5155
39	15,4	23,8	9,4	5668
34	13,4	21,0	8,3	6583
29	11,4	18,2	7,2	7080
24	9,5	15,6	6,1	8081
19	7,5	12,8	5,0	8709
14	5,5	9,6	3,8	10735
9	3,55	6,2	2,4	11869
0	0	0	0	15322

The points plotted in Figure 1. follow the equation (1) with the absorption coefficient $\mu = 0{,}044$/cm (0,112/inch).

In the winter season 1955-1956 an experimental snow-gaging station was built up in the Krkonoše mountains in the height of 1410 m (4600 feet). This time the radioactive source consisted of 30 mc Co 60. The Geiger-Müller tube was suspended on a construction 3m (9 feet 10 inches) above the soil surface. First of all the calibration

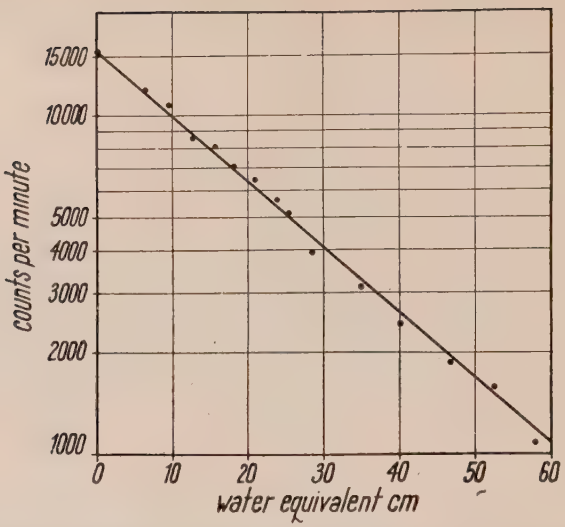

Fig. 1 — Snow-water-content measurements in the Krkonoše-mountains in 1955.

curve indicating the relation between the counting rate and the height of the water column was determined by introducing measured depths of water in a wooden pail installed between the radioactive source and the Geiger-Müller tube. The curve is presented in Figure 2.. For the highest water columns it was extrapolated. This time the absorption coefficient μ = 0,061/cm (0,154/inch) was calculated.

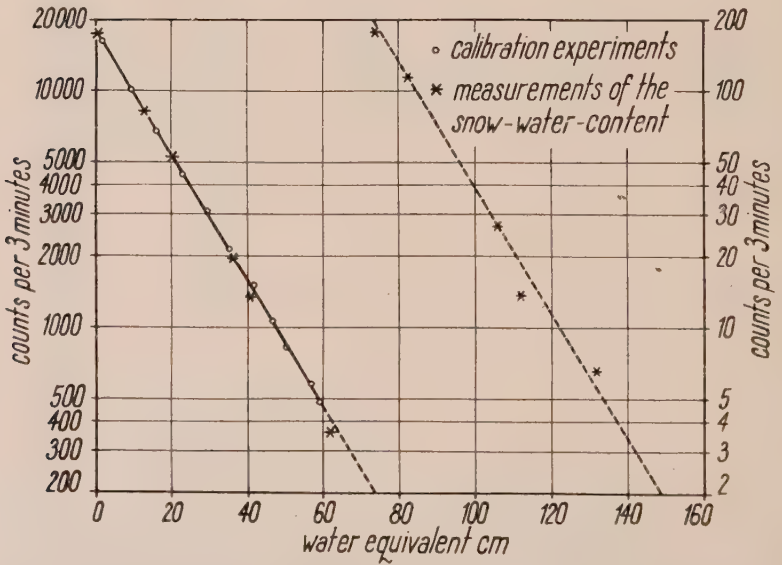

Fig. 2 — Snow-water-content measurements by melting and by radiocobalt in the Krkonoše mountains 1955-1956.

The measurements of the snow-water-content which were carried out from January 12 th until June 1st are presented in Table 2.

TABLE 2

Snow-water-content measurements by melting and by radiocobalt in the Krkonoše mountains 1955-1956

Date	Snow-depth		Water content		Distance correction a)	Decay corr.	Counting rate per 3 min. b)	Corrected counting rate per 3 min.
	cm	inch	cm	inch				
1955								
27.10	6,2	2,5	0,7	0,3	1,0	1,0	17116	17116
1956								
12.1.	210	82,9	74	29,1	1,0	1,025	175	179,4
14.1.	234	92,1	82,5	32,5	1,0	1,025	113	115,8
28.2.	245	96,5	106	41,7	0,7	1,04	36,6	26,6
28.3.	245	96,5	112	44,1	0,7	1,05	18,6	13,7
4.5.	235	92,5	132	52,0	0,97	1,06	6,45	6,6
30.5.	106	41,7	62	24,4	0,97	1,08	342,2	359,3
31.5.	69,5	27,4	41	16,1	0,97	1,08	1252,4	1315
1.6.	62	24,4	36	14,2	0,97	1,08	1854	1947
1.6.	35	13,8	20,3	8,0	0,97	1,08	5169,4	5428
1.6.	22	8,7	13	5,1	0,97	1,08	7863	8256

(To Table 2.)
a) the respective distances of the Geiger-Müller tube were 300 cm, 250 cm and 295 cm.
b) Cosmic background subtracted.

Before these results could be confronted with the calibration curve it was necessary to compensate for the decay of the radioctive source. The counting rates were also corrected with regard to some changes of the distance between the radioisotope and the Geiger-Müller tube, due to the deformation of the construction by snow. The correction coefficients are listed in Table 2. Further it was necessary to subtract from the total counts the cosmic background which could not be neglected as in the previous case in view of the reduced counting rates at the deepest snow packs. The comparative points which were obtained by sampling and melting the corresponding snow layers are plotted in Figure 2.

The unexpected heavy snowfalls in January and February caused the deformation of the construction which was snowed up for the greater part of the winter season. Of course, only the snow layer between the radioisotope and the Geiger-Müller tube was actually measured during this period.

In view of this experience a modified construction for holding the Geiger-Müller tube was designed by Ing. V. Sotornik. (Figure 3). Calibration measurements for this device were carried out in January 1957 in the Krkonoše mountains in the height of 850 m (2800 feet) with results presented in Table 3 and Figure 4.

Fig. 3 — Field installation radioactive snow-gage.

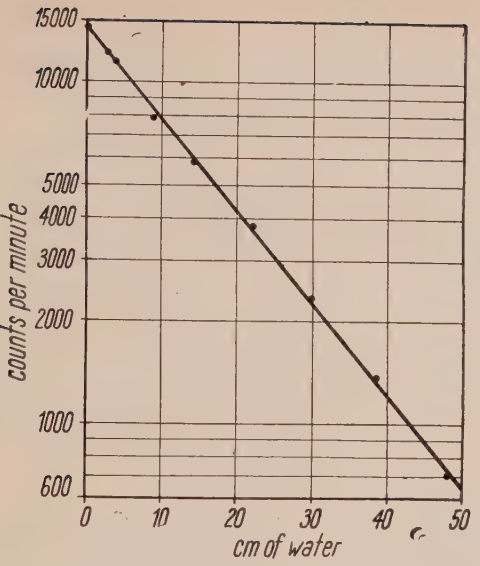

Fig. 4 — Calibration measurements in January 1957.

TABLE 3.

Calibration measurements in January 1957

Depth of water		Counting rate per minute cosmic backgr. subtracted	Depth of water		Counting rate per minute cosmic backgr. subtracted
cm	inch		cm	inch	
0	0	14411	21,7	8,5	3799
2,7	1,1	12035	29,7	11,7	2326
3,7	1,5	11391	38,5	15,1	1370
8,7	3,4	7783	48,1	18,9	707
14,2	5,6	5829			

The distance between the Geiger-Müller tube and the radioactive source (22 mc Co 60) was 2,5 m (8 feet 2 inches). Again a good agreement with the Eq. (1) was obtained and the absorption coefficient $\mu = 0{,}062$ /cm (0,157/inch) calculated.

Application of statistical theory to measurements of gamma radiation.

When the intensity of gamma radiation is measured by counting the pulses on a decimal scaler, the results of repeated measurements are somewhat different due to statistical fluctuations of the radiation. When the counting interval becomes

shorter and the measured intensity of radiation smaller, the relative deviations of the results become larger. The relation between the length of the counting interval and the accuracy of the result has been investigated in order to save time on routine measurements and to make the best use of the Geiger-Müller tube which has a limited life. During the operation of the experimental station in 1955-1956 a series of continuous measurements of the gamma radiation and of the cosmic background was carried out. During each experiment lasting one hour the partial increases of the total counts were noted every 10 seconds. Thus 360 values of partial intensities were obtained and treated by the statistical procedure which has been applied by Prochazka [2] for investigation of accuracy of water-velocity measurements.

This experimental evidence indicates that the partial increases of total counts are mutually independent and that they fluctuate about their mean value in accordance with the law of normal error. An axemple is presented in Figure 5.

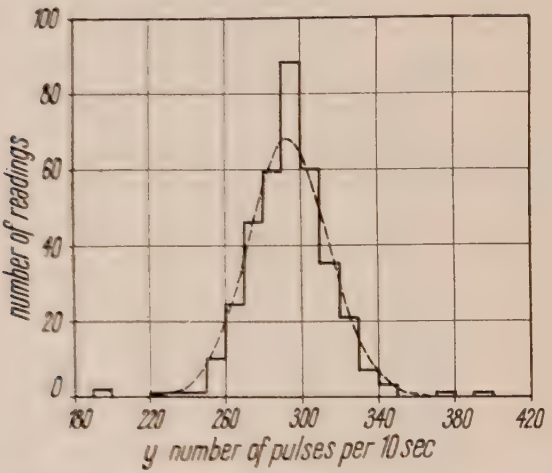

Fig. 5 — A comparison of the fluctuations of gamma radiation with the law of normal error.

Therefore we may write:

$$T = t\left(\frac{x \cdot C_v}{p}\right)^2 \qquad (2)$$

where T is the period of time over what a measurement must be made in order tha a desired accuracy of the result can be reached.

t is the partial time interval 10 sec.

x is the value of the normal-error function for a given probability

C_v is the coefficient of variation (the ratio of the standard deviation to the mean of the series).

p is the accuracy of the result.

From tables of the normal probability-integral

$$\Phi(x) = \frac{1}{\sqrt{2\pi}} \int_{-\lambda}^{+x} e^{-\frac{x^2}{2}} dx \qquad (3)$$

it is found that for the selected probability 0,90 the value of x is 1,645. This means that with the probability of 90 pct the given accuracy will be attained.

The values of C_v were investigated for various depths of snow and accordingly various intensities of gamma radiation. It was found that a relation existed which was expressed by the method of least squares:

$$C_v = 0{,}873\ \bar{y}^{-0{,}42} \tag{4}$$

where $\bar{y}$ is the mean number of pulses per 10 sec.

In the value of $\bar{y}$ the cosmic background is included. By eliminating the cosmic background the following equation for the gamma radiation alone was derived.

$$C'_v = \frac{1{,}294}{\sqrt{\bar{y}'}} \tag{5}$$

The intensity of fluctuation characterized by the coefficient of variation is inversely proportional to square root of the intensity of gamma radiation.

In a paper just published by Kursin and Michajlovskij (*) a similar relation was used in measuring the beta radiation.

For a given accuracy we obtain from Eqs. (3) and (4):

$$T = A \cdot \bar{y}^{0{,}84} \tag{6}$$

where A comprises all constants.

This relation between the intensity of gamma radiation (the cosmic background included) and the necessary duration of a single measurement for various values of desired accuracy is demonstrated in Fig. 6.

It should be remembered that Eqs. (4, 5, 6) were derived on the experimental station in 1955-1956 and need further verification in case they should be used for other arrangements of the measuring device.

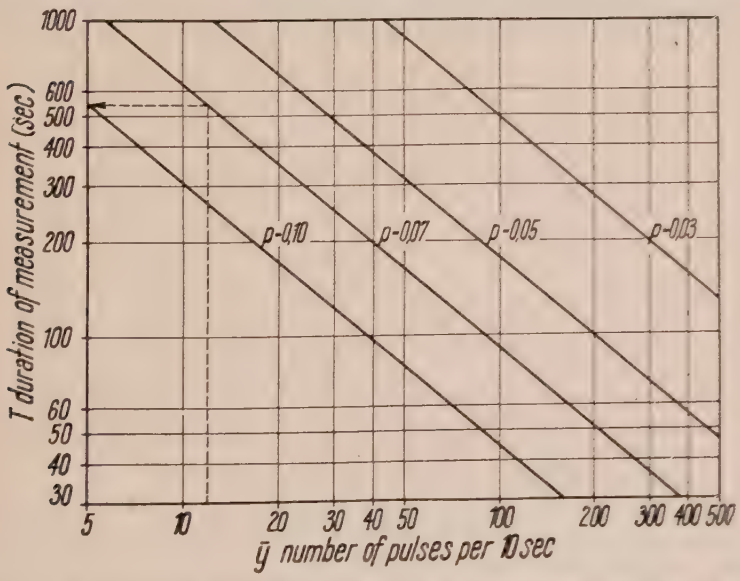

Fig. 6 — Relationships between the duration of measurement, and the intensity of gamma radiation for various values of desired accuracy.

Conclusions.

The results of experimental measurements presented in this paper indicate the suitability of the radioisotope Co 60 for measurements of the snow-water-content. Snow packs with the water equivalent up to 132 cm (52 inches) have been measured the absorption coefficient remaining a constant for all snow layers regardless of their depth (6 cm—245 cm, 2,5 inches—96,5 inches) and density (11 pct—60 pct).

The operation of an experimental snow-gaging station under unfavourable weather conditions in the winter season 1955-1956 provided the experience for the design of a reliable field equipment.

With this arrangement of the measuring device a close agreement of the absorption coefficient $\mu = 0{,}062$/cm (0,157/inch) with the theoretical value of 0,063/cm (0,160/inch) was reached.

The application of statistical theory made it possible to derive the relationships between the duration of a single measurement, the intensity of the gamma radiation and the accuracy of the result. By this procedure the duration of counts for routine measurements on a snow-gaging station can be economically determined.

As the pulses may be transmitted by radio, a radio communication system would be a suitable contribution to this method when applied for streamflow forecasting.

The radioactive snow gage may also be used on meteorologic stations for studies of snow properties without disturbance of the snow cover.

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FORECASTING THE SUMMER RUNOFF OF THE RHINE RIVER

J. E. CHURCH

SUMMARY

THE RHINE: The Rhine Basin comprises two distinct areas, viz., the Swiss or Upper Rhine and the Lower Rhine. The Swiss Rhine comprises only 22.5 percent of the entire basin but provides 43.0 percent of the annual runoff. On the basis of precipitation measurements, the efficiency of its annual runoff is 72.9 per cent as compared with 25.8 per cent on the Lower Rhine. This is even higher than the 67 per cent efficiency estimated for deep-snow runoff in the Central Sierra Nevada of North America. Switzerland is plainly a land of deep snow and shallow soil. The Lower Rhine, on the other hand, is an area of rain or quickly melting snow and deeper soil. Snow surveys would seem therefore to be restricted to the Upper Rhine, and the more stable forecasts would be confined there.

(*American Philosophical Society Year Book 1956*)

The Rhine River first came into prominence in streamflow forecasts in 1944 at the invasion of Germany during World War II but military needs almost immediately gave way to those of irrigation and power.

At the International Association of Hydrology at Oslo in 1948 Dr. Erwin Hoeck of Switzerland presented a plan for forecasting the runoff of the River Linth into Lake Zürich and Dr. J. Th. Thijsse of Netherlands pointed out the need of foreknowledge of the summer flow of the Rhine in order to maintain essential levels in the fresh-water Ijssel Lake for irrigating the reclaimed areas in the Zuider Zee.

Impelled by my recent snow-survey studies in the Himalaya and southern Andes, I visited Dr. Hoeck and other colleagues in Switzerland to analyze the character of the Rhine.

A close harmony in runoff was found to exist between the feeders of the Swiss Rhine and the Rhine at Basel and also the Rhone at Geneva. Furthermore it appeared that a snow survey even at the source of the Rhine above Lake Constance should be accurate within 15 percent for all of these streams during 4 years out of 5 with a maximum error of 35 percent in 21 years. The area involved was 100 × 150 miles but the snow cover probably represented the basic downstream summer flow. The basins studied were the Rhine above Lake Constance, the Linth above Lake Zürich, the Reuss at Lucerne, the Aar at Berne, and the combined Swiss Rhine at Basel or Rheinfelden.

Dr. Hoeck, who was Chief of the Hydrology Section of the Swiss Research Laboratory of Hydraulics and Earth Mechanics at Zürich undertook the task of developing a snow survey system for the Swiss Alps with the assistance of the Swiss Avalanche Laboratory at Weissfluh-Joch, Davos, which was conducting surveys of snow-texture in avalanche areas for the purpose of forecasting snow-slides.

The three-years results with comparisons based on the previous period beginning 1927 were prepared by Dr. Hoeck for presentation at the International Association in August at Brussels. The report bears the title, «La Prévision du débit estival de quelques rivières suisses par la méthode de corrélation à plusieurs variables» (The Forecasting of the Summer Delivery of Some Swiss Rivers by the Method of Correlation of Several Variables). But a fatal recurrence of cardiac embolism in the field at Chur in July prevented his expansion of the work. It must now be left to others.

The Main and the Moselle Rivers and minor neighboring tributaries below

Basel were left to me as my share in the cooperative plan. They are fed apparently by rain or quickly melting snow and their total flow in summer is less than that of the upper or Swiss Rhine. Eager cooperation has been afforded by Dr. Hoeck and by Dr. W. Friedrich of the Bundesanstalt für Gewässerkunde at Bielefeld, Germany and the Meteorological Service of France, including Prof. Maurice Pardé of the Institut de Géographie Alpine at Grenoble and Prof. J. P. Rothé of the Institut de Physique du Globe at Université de Strasbourg. My study, altho supplementary to the Swiss Rhine, must necessarily include all of the sources that ultimately merge in the Lowest Rhine. Moreover, because of the shortness or entire lack of snow survey records, dependence must be placed on precipitation records, with an accompanying lack of precision.

The Hydrology of the Rhine Basin

In streamflow efficiency or quantity of water supply, the Rhine Basin comprises two distinct areas, viz. the Swiss or Upper Rhine and the Lower Rhine. As shown in Table I, the Swiss Rhine comprises only 22.5 percent of the entire basin but provides 43.0 percent of the annual runoff. On the basis of precipitation measurements, the efficiency of its annual runoff is 72.9 percent as compared with 25.8 percent on the Lower Rhine. This is even higher than the 67 percent efficiency estimated for deep-snow runoff in the Central Sierra Nevada of North America. Switzerland is plainly a land of deep snow and shallow soil. The Lower Rhine, on the other hand, is an area of rain or quickly melting snow and deeper soil. Snow surveys would seem therefore to be restricted to the Upper Rhine, and the more stable forecasts would be confined there.

Complexities appear, moreover, thruout the basin. On the basis of 3 precipitation stations on the Moselle-Meurthe 1 on the Main, and 6 in Switzerland (Table 2), the heaviest annual precipitation is on the Moselle-Meurthe 1775 mm (69.2 in.) and the lowest is on the Main 918 mm (35.8 in.). In Switzerland the average is 1236 mm (48.2 in.). The heaviest precipitation is 1819 mm at Eigergletscher at 2323 m (7600 ft.) above sea level but this is slightly exceeded in the Moselle by 2411 mm at Lac d'Alfeld at the comparatively low elevation of 620 m (2000 ft.) in the French Vosges.

The distribution of precipitation also is not conducive to forecasting streamflow tho downstream its evenness does assure continuous power. In Switzerland the precipitation is heaviest in midsummer and increases from 37.6 percent of annual during the 6 winter months of October-March to 62.4 percent during the summer months of April-September, the period for which exact forecasts are desired. Below Switzerland, the precipitation is nearly uniform thruout the year.

The runoff pattern also follows the precipitation pattern closely even in the Swiss or snow sector, as seen in the following excerpt based on percentage of annual from Table 1.

Sector	Precipitation		Runoff	
	Winter	Summer	Winter	Summer
Swiss	37.6	62.4	37.2	62.8
Main	49.8	49.8	61.3	38.8
Moselle-Meurthe	54.7	45.3	56.1	44.0
Total Rhine	48.8	51.1	52.8	47.2

Divisions	Area Km²	Pct. of Total		Elevation of Stations (m above sea)	Mean Precipitation Winter (mm)	(Oct.-Mar.) (% of Ann.)	Summer (mm)	(Apr.-Sept.) (% of Ann.)	Annual (mm)	(In.)
1. Upstream										
Swiss Rhine (Basel)	35,925	22.5		453 — 2323	468	37.6	767	62.4	1236	48.2
2. Downstream										
Main (Schweinfurt)	12,719	8.0		472	457	49.8	461	49.8	918	35.8
Moselle-Meurthe (Kochem)	28,156	17.6	77.5	370 — 620	993	54.7	782	45.3	1775	69.2
Other-Sources	82,883	51.9			725	53.9	622	46.2	1346	52.5
123,758										
3. Entire Rhine (Rees)	159,683	100.0			639	48.8	670	51.1	1310	51.1

TABLE 1 (*continued*)

Divisions	Main Runoff Winter (m³/s)	Oct.-Mar.) (% of Ann.)	Summer (m³/s)	Apr.-Sept.) (% of Ann.)	(m³/s)	Efficiency of Runoff Annual (% of Entire stream)	Annual Pct.	Only of Precip.
1. Upstream								
Swiss Rhine (Basel)	763	37.2	1290	62.8	1027	43.0	72.9	Oct.-Mar. 67.7% Apr.-Sept. 67.7%
2. Downstream								
Main (Schweinfurt)	135	61.3	86	38.8	110	4.6	29.7	
Moselle-Meurthe (Kochem)	425	56.1	332	44.0	384	16.1 57.0	24.2	25.8
Other Sources	1307	66.8	644	33.0	869	36.4	24.6	
3. Entire Rhine (Rees)	2630	52.8	2352	47.2	2390	100.0	36.0	
					(= 61,064,500 AF)			

TABLE 2

Precipitation Chart of Rhine River Basin

Sub-Basins, Stations	Mean Monthly Precipitation (mm)														
	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Total	Apr.	May	Je	Jly.	Aug.	Sept.	Total	Annual
1. Upper Rhine R. (35,925 Km²)															
Swiss Alps (Only 22.5% of total)															
Aar R. Berne (alt. 572 m.)	76	64	51	53	44	64	352	68	101	127	132	108	98	634	986
Reuss R. Lucerne (alt. 453 m)	81	63	55	64	56	67	386	79	124	148	158	152	107	768	1154
Linth R. Glarns (alt. 480 m)	109	103	86	101	93	94	586	95	129	175	199	178	123	899	1485
Rhine R. Chus (alt. 610 m)	69	62	46	46	52	56	331	51	77	93	116	106	78	521	852
Eigergletscher (alt. 2323 m)	134	123	109	117	117	133	733	155	200	191	207	185	148	1086	1819
Safieuplatz (alt. 1270 m)	102	81	54	58	60	67	422	76	113	124	145	135	103	696	1118
							468							767	1236 (= 48.2 inch.)
2. Lower Rhine R.															
Main R. (12,719 Km²)															
Erz Gebirge															
Johanneskreut (alt. 472 m)	80	85	81	79	62	70	457	72	68	76	85	85	75	461	918 (= 35.8 in.)
Speyerbach															
Moselle-Meurthe R. (28,156 Km²)															
Hautes Vosges															
St. Michel (alt. 370 m)	123	103	81	101	59	82	579	114	90	120	116	105	108	653	1232
Ramonchamp (474 m)	159	187	161	220	135	114	976	111	109	134	129	114	110	707	1683
Lac d'Alfeld (620 m)	220	278	229	290	222	186	1425	174	146	176	169	157	164	986	2411
							993							782	1775 (= 69.2 in.)
Other Downstream Areas (82,883 Km²)	No data														
Total Lower R. 123,758 Km² (77.5% of Basin)	Estimate on basis of Main and Moselle-Meurthe						725							622	1346 (= 52.5 in.)

Sub-Basins, Stations	Mean-Monthly Percentage of Annual													
	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Total	Apr.	May	Je.	Jly.	Aug.	Sept.	Total
1. Upper Rhine R. (35,925 Km²)														
Swiss Alps (Only 22.5% of total)														
Aar R. Berne (alt. 572 m)	7.7	6.5	5.2	5.4	4.5	6.5	35.8	6.9	10.2	12.9	13.4	10.9	9.9	64.2
Reuss R. Lucerne (alt. 453 m)	7.0	5.5	4.8	5.5	4.9	5.8	33.5	6.8	10.7	12.8	13.7	13.2	9.3	66.5
Linth R. Glarns (alt. 480 m)	7.3	6.9	5.8	6.8	6.3	6.3	39.4	6.4	8.7	11.8	13.4	12.0	8.3	60.6
Rhine R. Chus (alt. 610 m)	8.1	7.3	5.4	5.4	6.1	6.6	38.9	6.0	9.0	10.9	13.6	12.4	9.1	61.0
Eigergletscher (alt. 2323 m)	7.4	6.8	6.0	6.4	6.4	7.3	40.3	8.5	11.0	10.5	11.4	10.2	8.1	59.7
Safieuplatz (alt. 1270 m)	9.1	7.2	4.8	5.2	5.4	6.0	37.7	6.8	10.1	11.1	13.0	12.1	9.2	62.3
	7.8	6.7	5.3	5.8	5.6	6.4	37.6	6.9	10.0	11.7	13.1	11.8	9.0	62.4
2. Lower Rhine R.														
Main R. (12,719 Km²)														
Erz Gebirge														
Johanneskreut (alt. 472 m)	8.7	9.3	8.8	8.6	6.8	7.6	49.8	7.8	7.4	8.3	9.3	9.3	7.7	49.8
Speyerbach														
Moselle-Meurthe R. (28,156 Km²)														
Hautes Vosges														
St. Michel (alt. 370 m)	10	8.4	6.6	8.2	7.2	6.7	47.1	9.3	7.3	9.7	9.4	8.5	8.8	53.0
Ramonchamp (474 m)	9.4	11.1	9.6	13.1	8.0	6.8	58.0	6.6	6.5	8.0	7.6	6.8	6.5	42.0
Lac d'Alfeld (620 m)	9.1	11.5	9.5	12.0	9.2	7.7	59.0	7.2	6.0	7.3	7.0	6.5	6.8	40.8
	9.5	10.3	8.6	11.1	8.1	7.1	54.7	7.7	6.6	8.3	8.0	7.3	7.4	45.3
Other Downstream Areas														
(82,883 Km²)														
Total Lower R. 123,758 Km²														
(77.5% of Basin)							53.9							46.2
3. Entire Rhine Basin							46.2							53.8
(159,683 Km² to Rees)														

TABLE 3

Runoff of the Rhine River Basin

Sub-Basins, Stations	Mean Monthly Runoff (m^3/s)														
	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Total	Apr.	May	June	July	Aug.	Sept.	Total	Annual
UPPER RHINE R.															
Total at Basel (35,925 Km^2 1808-1947)	921	813	739	669	664	774	763	975	1243	1531	1523	1342	1125	1290	1027
Rhine Source															
Constance															
Zürich															
Lucerne															
Neuchatel															
LOWER RHINE R.															
Main, Schweinfurt (12,719 Km^2) (1925-1947)	79	125	114	145	173	174	135	140	85	81	75	71	61	86	110
Moselle-Meurthe, Kochem (28,156 Km^2) (1937-1948)	386	465	366	372	534	427	425	322	331	316	385	389	252	332	379
Other Sources (82,883 Km^2)	904	927	1221	1574	1629	1585	1307	1413	621	352	427	458	592	644	975
Total	1369	1517	1701	2091	2336	2186	1867	1875	1037	749	887	918	905	1062	1464
ENTIRE RHINE R.															
Total at Rees (159,683 Km^2) (1936-1940)	2290	2330	2440	2760	3000	2960	2630	2850	2280	2280	2410	2260	2030	2352	2390

TABLE 3 (*continued*)

Runoff of the Rhine River Basin

Sub-Basins, Stations	Mean Monthly Percentage of Annual													
UPPER RHINE R.	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Total	Apr.	May	June	July	Aug.	Sept.	Total
Total at Basel (35,925 Km² 1808-1947)	7.5	6.6	6.0	5.4	5.4	6.3	37.2	7.9	10.1	12.4	12.4	10.9	9.1	62.8
Rhine Source														
Constance														
Zürich														
Lucerne														
Neuchatel														
LOWER RHINE R.														
Main, Schweinfurt (12,719 Km²) (1925-1947)	6.0	9.4	8.6	11.0	13.1	13.2	61.3	10.6	6.4	6.1	5.7	5.4	4.6	38.8
Moselle-Meurthe, Kochem (28,156 Km²) (1937-1948)	8.5	10.2	8.1	8.2	11.7	9.4	56.1	7.1	7.3	7.0	8.5	8.6	5.5	44.0
Other Sources (82,883 Km²)	7.7	7.9	10.4	13.4	13.9	13.5	66.8	12.1	5.3	3.0	3.6	3.9	5.1	33.0
Total	7.8	8.6	9.7	11.9	13.3	12.4	63.7	10.7	5.9	4.3	5.0	5.2	5.2	36.3
ENTIRE RHINE R.														
Total at Rees (159,683 Km²) (1936-1940)	7.7	7.8	8.2	9.2	10.0	9.9	52.8	9.5	7.6	7.6	8.1	7.6	6.8	47.2

The monthly runoff seems also to parallel the precipitation (Tables 2 and 3) except on the Main, where runoff exceeds precipitation in January thru April and falls below it the remainder of the summer. The parallelism seems remarkable in Switzerland in the presence of the obvious accumulation of snow in the winter and melting in the summer. The open-water area that would melt and deliver the snow as it falls should, however, scarcely exceed 4 percent of the basin. On the other hand, retention by dams seems slight.

Exact figures are not available. The great lakes of Constance, Zürich, Lucerne, and Neuchatel cover 375.75 sq. mi. Thun, Brienz, Wallen, Sempach cover 5 ½ sq. mi. each. No others exceed 4 sq. mi. This might make a total of 500 sq. mi. in 14,000 sq. mi. of basin.

The Basis of Snow Surveying

The unique value of Snow Surveying as compared with accumulated measurements of intermittent precipitation lies in the net efficiency of the snow cover as a source of water and the relative ease of measuring its potential at the end of the season of accumulation and beginning of seasonal flow.

This measurement in terms of water equivalent of a few fixed courses in each watershed is reduced to seasonal percentage and correlated with the runoff of the stream at the base of the snow cover. This has been called the percentage or Nevada method in token of the place of its development.

Close correlation of snow cover and runoff is rendered easy by the relatively stable winter precipitation over the watershed. For example, in the Truckee River Basin of California-Nevada during 25 years a single snow course at Donner Pass at the head of the basin indicated the snow-melt runoff (April-July) 14 times within 10 percent of actual. Furthermore, the snow cover over several adjacent basins maintains a seasonal uniformity within 25 percent of each other for 100 miles, with corresponding harmony of streamflow. The basins of Switzerland have been found even more harmonious.

Correction Factors

With the percentage system variables are few and most of them can be determined at the time of forecast. These are (*a*) premature melting of the lower zone of the snow cover, (*b*) effect of diversions above point of streamgaging, (*c*) relative loss or even gain in priming the soil or meadows along the stream channel.

Premature melting can be determined closely by altitude zones for the snow surveys. Soil priming is a problem chiefly where the soil is alluvial and autumn rains are heavy. On steeper mountain slopes the annual loss from priming is nearly uniform and its percentage effect is dependent, like diversions, on the quantity of the snow cover.

The only unpredictable factor is the precipitation during snow-melt. Close correlation requires normal precipitation on the snow while it is in process of melting and its lack or excess will distort the snow survey percentage by 10 to 15 percent of normal, or even more depending upon the relative precipitation during this period. The precipitation during April or April-May (In the Southern Hemisphere October-November) may provide data for a revised forecast.

Temperature is a factor in the acceleration and retardation of snow-melt but not in its total volume, which depends on the water equivalent of the snow cover. Only rarely does temperature departure seem to affect the total amount forecasted. This stability is due to the normal freezing of the snow cover each night with diurnal

cessation of melting and the protective thermal zone at the snow surface that retards the effect of the sun's heat above it. Thus on May 6, 1949 a snow crust 6-9 inches deep was formed at Soda Springs with a minimum temperature in the vicinity of 37° F. 14 feet above the ground. But at 7:30 a.m. the surface of the snow was 32.5°F. although at 1 inch above it the temperature was 38°F. Similarly, at 2:15 p.m. April 16, at Truckee, the following shade temperatures were recorded: 9 feet 68.5°F., 18 inches 67.0°F., on snow 41°F.
These stations were at 6,750 and 6,000 feet altitude, respectively, in the Sierra Nevada in California.

Efficiency in Snow Cover

The efficiency of the snow cover as compared with intermittent rain is due to the fact that the snow cover provides a «continuous» rain that requires only one initial priming of the soil while each of the intermittent rains must provide its own priming in proportion to the dryness of the soil. This is naturally less true of intermittent snow that melts instead of accumulating.

For example, in Hawaii (Monthly Weather Review 45: 178-81) the runoff in volcanic soil and verdure varied in rains of ½ inch from 7.3 percent efficiency after a period of very dry weather to 30.0 percent after heavy precipitation and with rains of 2 inches from 18.6 percent after very light showers to 34.6 percent after heavy precipitation. In sand-box studies by the writer the even flow of 29, 69, and 83 gallons of water for the period of 11, 17, and 23 days, respectively, had an efficiency of 60.6, 66.6, and 63.6 percent or approximately the observed efficiency of the deep snow cover in the South Yuba Basin of the Central Sierra Nevada and in Switzerland. Thus in net efficiency a snow cover has twice the potential of a similar quantity of rain distributed over an equal period. Thus in the Susquehanna Basin of eastern U.S.A. the snow cover constitutes only one third of the annual precipitation but is far more effective in runoff than the remaining two thirds.

This greater efficiency of the snow cover, as compared with rain, renders the snow cover the dominant factor in streamflow irrespective of the precipitation in the summer season. Thus the phases of the Sierra Nevada streams do not vary greatly from those of the Continental Divide, although the summmer precipitation (Apr.-July) varies from 18.2 percent of the winter (Nov.-Mar.) in the Central Sierra Nevada (Colfax) to 85.7 percent in the Continental Divide (Durango, Telluride, Leadville) and 126.5 and 284.9 percent farther east (Banff and Calgary). The normal phases of typical streams are as follows:

Uniformity of Snow-melt Runoff Irrespective of Summer Rainfall

Seasonal Phases	(Percentage of Annual) Sierra Nevada San Joaquin (Ann. 2,053,500 A.F.)	West slope of Continental Divide, Upper Colorado (Ann. 7,540,000 A.F.)	East Slope of Continental Divide Bow (Ann. 2,704,150 A.F.)
December-March	21.3	9.5	Dec.-Apr. 13.2
Apr.-July (Snow-melt)	70.1	75.0	May-Aug.* 68.4
August-November	8.6	15.5	Sep.-Nov. 18.4

*Season retarded by latitude.

68 to 75 percent of the annual normal runoff is due to the snow cover of Dec.-March or Dec.-April and the subsequent summer precipitation that falls upon it. Therefore, it has been possible during 11 years of test to forecast the April-July runoff of the Colorado River on May 1 within 15 percent.

Since the August-November runoff in the Sierra Nevada is still smaller, closer forecasts can be made of its streams. The presence of sagebrush on the foothills of the Continental Divide and Sierra Nevada indicates lack of efficient summer rains. Only farther eastward, where the measured rain is much heavier, do the grassland and rainfed streams appear.

Defects in Snow Gage Measurements

All methods of streamflow forecasting based on falling snow instead of snowcover residue are handicapped by inaccurate measurements due to wind and snow caps but particularly to the fact that the can does not measure the wastage or premature melting of the snow, that is, the net amount of snow-melt water to be expected.

Anomalies on the Rhine

The scarcity of snow-survey data on the Upper or Swiss Rhine and of snowfields on the Lower Rhine has required the use of less efficient precipitation measurements for purpose of analysis. However, the data in Tables 1-2, being longer averages rather than annuals, should produce an accurate picture.

On the Lower Rhine, where rain predominates, parallelism between monthly precipitation and runoff is expected but on the Upper Rhine the parallelism is even closer in both winter and summer despite the fact that snow, accumulated largely in winter, is recognized as the base of the summer flow.

The quick transport of water at all seasons by the shallow soil, the transfer of snow to water in winter by open, free-flowing lakes and rivers and the presence of major precipitation in summer and its maximum intensity in July quite effectively conceal the snow-melt cycle: They even almost conceal its lengthened period of 6 months (April-September) due to the higher mean elevation of the Alps and retardation of flow by lakes.

Correction Factors

Soil Priming — On all feeders of the Rhine (Table 3) the minimum pre-winter flow is scarcely less than one-half of the maximum seasonal flow and, therefore, the soil should remain continuously moist as compared with semi-arid climates and dry summers. On the other hand, the soil of the Upper Rhine is shallow and drains quickly. Only in extreme dryness, would the correction factor be necessary. Ground-water measurements should be used to supplement the surface flow as an index of soil moisture.

In the absence of impounding dams, the level of the lakes on April 1 and subsequent rise should be supplemental to measurements of streamflow. On large lakes, such as Lake Tahoe, California-Nevada, the forecast is made in percentage of normal or feet rise from April 1, to maximum.

The great efficiency of the soil, snow and water mantle on the Upper Rhine is evidenced by the net delivery of 67.7 percent of precipitation in October-March and the same percentage also in April-September. This must include also the evaporation loss during winter and summer. The low efficiency of intermittent rains on deeper soil is shown by the low efficiency of 25.8 percent on the Lower Rhine.

Summer Rains. — Summer rains and snows are more efficient when they fall on snow, ice or water surfaces and lose little of their volume from dissipation by the soil. Likewise their efficiency diminishes with decrease in the area of the snow fields. Thus at the beginning of the snow-melt season, as in April, the efficiency of precipitation should be 100 percent and shrink to probably 25 percent when the snow has disappeared. This has largely been accomplished by late July.

On the other hand, the summer precipitation attains its maximum of 13.1 percent of annual in July (Table 2) while maintaining a volume of 10.0 to 11.3 percent during May-August. The normal steadiness of precipitation assures a fair approximation to normal. For this reason Dr. Hoeck has selected April-July (the usual 4 month period) for his forecasts of runoff.

On the 6-month basis, the summer precipitation in Switzerland of 767 mm is 163.9 percent of the winter precipitation of 468 mm. But if the 4-month period of April-July is used, its total of 514 mm would be 109.8 percent of the winter or only 24 percent higher than on the American Continental Divide. There the correction as computed on May 1 is 15 percent for extreme excess or deficiency in summer precipitation with no large lakes in the watershed. In Switzerland, Dr. Hoeck's forecasts (Table 7) of the highest feeders came within 15 percent of measured flow in 90 percent of the 24 years tested. This was based on measurements of precipitation mostly.

Phases of Snow-Melt. — The general phases of snow-melt would seem to parallel closely those in Western North America from which comparative data are being drawn, if the following temperatures in the two areas can be taken as a criterion

	Mean Temperature (°F)	
Low Stations	January	July
Basel (909 ft.)	31.8	66.4
Reno (4,500 ft.)	30.9	68.1
Denver (5,299 ft.)	30.4	72.2
High Stations		
Davos (5,121 ft.)	18.7	53.8
Truckee (6,000 ft.)	25.7	63.4

On Säntis the precipitation is snow from November thru April, and only in July-August is there more rain than snow.

Forecasting the Flow of the Upper Rhine: A Quarter Century Test.

For lack of snow-survey data, the longer series of precipitation measurements has been used to compare snow-melt or even precipitation with subsequent runoff. Direct measurements and percentage of normal have been used for each year since 1925-26. The winter (Oct.-Mar.), summer (Apr.-Sept.) and annual phases have been computed (Table 3, 4, 5). Monthly data could afford more intimate clues to interrelationships particularly of summer precipitation to summer flow.

Table 6 — Precipitation and Runoff on Upper Rhine (April-September) was limited to the years from 1935-36 thru current lack of data. It presents the unweighted percentages of Snow Cover (Linth or Zürich Basin) April 1, Winter Precipitation, Winter Runoff and Summer Runoff, followed by Summer Precipitation to seek correlations.

TABLE 4

Annual Precipitation on the Upper Rhine
(mm and percent of Normal)

	Berne (alt. 572M)						Lu;erne (alt. 453 M)					
	MM			Percent of Normal			MM			Percent of Normal		
	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual
1925-26	402.0	590.0	992.0	113.8	93.1	100.3	317.0	804.0	1121.0	81.2	104.8	96.8
1926-27	459.0	793.0	1252.0	129.9	125.1	126.6	440.0	983.0	1423.0	112.7	128.1	122.9
1927-28	323.0	581.0	904.0	91.4	91.7	91.4	278.0	682.0	960.0	71.2	88.9	82.9
1928-29	300.0	466.0	766.0	84.9	73.5	77.4	339.0	616.0	955.0	86.8	80.3	82.5
1929-30	427.0	838.0	1265.0	120.9	132.2	127.9	443.0	930.0	1373.0	113.4	121.2	118.6
1930-31	612.0	736.0	1348.0	173.2	116.1	136.3	648.0	772.0	1420.0	165.9	100.6	122.6
1931-32	238.0	696.0	934.0	67.4	109.8	94.4	296.0	808.0	1104.0	75.8	105.3	95.3
1932-33	277.0	571.0	848.0	78.4	90.1	85.7	311.0	763.0	1074.0	79.6	99.4	92.7
1933-34	255.0	534.0	789.0	72.2	84.2	79.8	345.0	669.0	1014.0	88.3	87.2	87.6
1934-35	431.0	651.0	1082.0	122.0	102.7	109.4	367.0	723.0	1090.0	94.0	94.2	94.1
1935-36	563.0	838.0	1401.0	159.3	132.2	141.6	556.0	957.0	1513.0	142.4	124.7	130.7
1936-37	508.0	640.0	1148.0	143.8	101.0	116.1	422.0	735.0	1157.0	108.1	95.8	99.9
1937-38	249.0	735.0	984.0	70.5	115.9	99.5	349.0	799.0	1148.0	89.4	104.1	99.1
1938-39	282.0	679.0	961.0	79.8	107.1	97.2	304.0	772.0	1076.0	77.8	100.6	92.9
1939-40	428.0	761.0	1189.0	121.1	120.0	120.2	605.0	976.0	1581.0	154.9	127.2	136.5
1940-41	415.0	544.0	959.0	117.5	85.8	97.0	388.0	759.0	1147.0	99.4	98.9	99.0
1941-42	221.0	568.0	789.0	62.6	89.6	79.8	339.0	620.0	959.0	86.8	80.8	82.8
1942-43	246.0	582.0	828.0	69.6	91.8	83.7	326.0	611.0	937.0	83.5	79.6	80.9
1943-44	169.0	581.0	750.0	47.8	91.7	75.8	321.0	760.0	1081.0	82.2	99.0	93.4
1944-45	458.0	513.0	971.0	129.6	80.9	98.2	497.0	798.0	1295.0	127.3	104.0	111.8
1945-46	240.0	749.0	989.0	67.9	118.2	100.0	291.0	927.0	1218.0	74.5	120.8	105.2
1946-47	318.0	279.0	597.0	90.0	44.0	53.3	323.0	470.0	793.0	82.7	61.2	68.5
1947-48	350.0	654.0	1004.0	99.1	103.2	101.5	477.0	718.0	1195.0	122.7	93.6	103.2
1948-49												
1949-50												
Mean	355.3	633.9	989.1				390.5	767.5	1158.0			

Annual Precipitation on the Upper Rhine
(mm and percent of Normal)

	Glarus (alt. 480m)						Chur (alt. 610m)					
	mm			Percent of normal			mm			Percent of Normal		
	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual	Nov.-Mar.	Apr.-Sept.	Annual	Nov.-Mar.	Apr.-Sept.	Annua
1925-26	574.0	918.0	1492.0	96.8	102.5	100.2	311.0	553.0	864.0	92.4	106.4	100.9
1926-27	729.0	1208.0	1937.0	122.9	134.9	130.1	483.0	716.0	1199.0	143.5	137.8	140.0
1927-28	386.0	840.0	1226.0	65.1	93.8	82.4	229.0	635.0	864.0	68.1	122.2	100.9
1928-29	453.0	797.0	1250.0	76.4	89.0	84.0	365.0	449.0	814.0	108.5	86.4	95.1
1929-30	442.0	1081.0	1523.0	74.5	120.7	102.3	235.0	574.0	809.0	69.8	110.5	94.5
1930-31	791.0	864.0	1655.0	133.4	96.5	111.2	384.0	592.0	976.0	114.1	113.9	114.0
1931-32	350.0	839.0	1189.0	59.0	93.7	79.9	240.0	548.0	788.0	71.3	105.5	92.0
1932-33	460.0	1092.0	1552.0	77.6	121.9	104.3	260.0	531.0	791.0	77.3	102.2	92.4
1933-34	366.0	852.0	1218.0	61.7	95.1	81.8	256.0	540.0	796.0	76.1	103.9	93.0
1934-35	724.0	846.0	1570 .0	122.1	94.5	105.5	476.0	459.0	935.0	141.5	88.3	109.2
1935-36	762.0	1025.0	1787.0	128.5	114.5	120.0	485.0	640.0	1125.0	144.1	123.2	131.4
1936-37	691.0	1038.0	1729.0	116.5	115.9	116.2	409.0	412.0	821.0	121.5	79.3	95.9
1937-38	492.0	814.0	1306.0	83.0	90.9	87.7	298.0	404.0	702.0	88.6	77.8	82.0
1938-39	466.0	918.0	1384.0	78.6	102.5	93.0	270.0	539.0	809.0	80.2	103.8	94.5
1939-40	908.0	1235. 0	2143.0	153.1	137.9	144.0	441.0	559.0	1000.0	131.1	107.6	116.8
1940-41	536.0	758.0	1294.0	90.4	84.6	86.9	295.0	431.0	727.0	88.0	82.9	84.9
1941-42	560.0	752.0	1312.0	94.4	84.0	88.1	231.0	502.0	733.0	68.6	96.6	85.6
1942-43	438.0	778.0	1216.0	73.9	86.9	81.7	207.0	543.0	750.0	61.5	104.5	87,6
1943-44	578.0	678.0	1256.0	97.5	75.7	84.4	221.0	338.0	559.0	65.7	65.0	65.3
1944-45	1003.0	961.0	1964.0	169.1	107.3	131.9	530.0	507.0	1037.0	157.5	97.6	121.1
1945-46	565.0	790.0	1355.0	95.3	88.2	91.0	365.0	516.0	881.0	108.5	99.3	102.9
1946-47	462.0	608.0	1070.0	77.9	67.9	71.9	250.0	435.0	635.0	74.3	83.7	80.0
1947-48	902.0	905.0	1907.0	152.1	101.1	121.4	498.0	528.0	1026.0	148.0	101.6	119.8
1948-49												
1949-50												
Mean	593.0	895.5	1488.5				336.5	519.6	856.1			

TABLE 4 (*continued*)

Annual Precipitation on the Upper Rhine
(mm and percent of Normal)

	Eigergletscher in Aaregebiet						Safienplatz (alt. ca. 1270m)					
	mm			Percent of Annual			mm			Percent of Normal		
	Nov.-Mar	Apr.-Sept.	Annual	Nov.-Mar.	Apr.-Sept.	Annual	Nov.-Mar.	Apr.-Sept.	Annual	Nov.-Mar.	Apr.-Sept.	Annual
1925-26	671.0	1310.0	1981.0	90.5	120.6	108.4	366.0	678.0	1044.0	85.5	97.3	92.8
1926-27	1021.0	1126.0	2147.0	137.7	103.7	117.5	692.0	870.0	1562.0	161.7	124.5	138.9
1927-28	517.0	1122.0	1639.0	69.7	103.3	89.7	334.0	727.0	1061.0	78.1	104.3	94.3
1928-29	942.0	991.0	1933.0	127.1	91.3	105.8	530.0	582.0	1112.0	123.9	83.5	98.9
1929-30	627.0	1498.0	2125.0	84.6	138.0	116.3	350.0	748.0	1098.0	81.8	107.3	97.6
1930-31	1114.0	1104.0	2218.0	150.3	101.7	121.4	485.0	762.0	1247.0	113.3	109.4	110.9
1931-32	372.0	1144.0	1516.0	50.2	105.4	83.0	256.0	790.0	1046.0	59.8	113.4	93.0
1932-33	555.0	1082.0	1637.0	74.9	99.6	89.6	347.0	732.0	1079.0	81.1	105.1	95.9
1933-34	730.0	1011.0	1741.0	98.5	93.1	95.3	366.0	730.0	1096.0	85.5	104.8	97.4
1934-35	821.0	1012.0	1833.0	110.8	93.2	100.3	518.0	597.0	1115.0	121.1	85.7	99.1
1935-36	979.0	1259.0	2238.0	132.1	116.0	122.5	659.0	852.0	1511.0	154.0	122.3	134.3
1936-37	619.0	1166.0	1785.0	83.5	107.4	97.7	482.0	656.0	1138.0	112.6	94.1	101.2
1937-38	568.0	1092.0	1660.0	76.6	100.6	90.9	367.0	600.0	967.0	85.8	86.1	86.0
1938-39	575.0	1229.0	1804.0	77.6	113.2	98.7	346.0	819.0	1165.0	80.9	117.5	103.6
1939-40	909.0	1187.0	2096.0	122.6	109.3	114.7	561.0	686.0	1247.0	131.1	98.4	110.9
1940-41	670.0	1263.0	1933.0	90.4	116.3	105.8	387.0	587.0	974.0	90.4	84.2	86.6
1941-42	586.0	851.0	1437.0	79.0	78.4	78.6	289.0	676.0	965.0	67.5	97.0	85.8
1942-43	644.0	981.0	1625.0	86.9	90.3	88.9	329.0	742.0	1071.0	76.9	106.5	95.2
1943-44	702.0	904.0	1606.0	94.7	83.3	87.9	285.0	538.0	823.0	66.6	77.2	73.2
1944-45	1103.0	1050.0	2153.0	148.8	96.7	117.8	589.0	662.0	1251.0	137.6	95.0	111.2
1945-46	752.0	778.0	1530.0	101.4	71.7	83.7	492.0	707.0	1199.0	115.0	101.5	106.6
1946-47	655.0	727.0	1382.0	88.4	87.0	75.6	311.0	651.0	962.0	72.7	93.4	85.5
1947-48	918.0	1086.0	2004.0	123.8	100.0	109.7	500.0	635.0	1135.0	116.8	91.1	100.9
1948-49												
1949-50												
Mean	741.3	1085.8	1827.1				427.9	696.8	1124.7			

TABLE 4 (*continued*)

Annual Precipitation on the Upper Rhine
(mm and percent of Normal)

	Oct.-Mar.	Average Based on Mean of 1925-48 Apr.-Sept.	Annual
1925-26	93.4	104.1	99.4
1926-27	134.7	125.7	129.3
1927-28	73.9	100.7	90.3
1928-29	101.3	84.0	90.6
1929-30	90.8	121.6	109.5
1930-31	141.7	106.4	119.4
1931-32	63.9	105.5	89.6
1932-33	78.2	103.0	93.4
1933-34	80.4	94.7	89.2
1934-35	118.6	93.1	102.9
1935-36	143.4*	122.2	130.1
1936-37	114.3	98.9	104.5
1937-38	82.3	95.9	90.9
1938-39	79.2	107.4	96.6
1939-40	135.6	116.7	122.4
1940-41	96.0	92.1	93.4
1941-42	76.5	87.7	83.4
1942-43	75.4	93.3	86.3
1943-44	75.8	82.0	80.0
1944-45	145.0	96.9	117.0
1945-46	93.8	100.0	98.2
1946-47	81.0	69.5	72.5
1947-48	127.1	98.4	109.4
1948-49			
1949-50			
Mean	*Normal Based on 1935-1948		

TABLE 5

Annual Precipitation on Lower Rhine

(mm and Percent of Normal)

	St. Michel (alt. 370m)						Ramonchamp (alt. 474m)					
	mm			Percent of Normal			mm			Percent of Normal		
Normal	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual
1925-26	627.8	668.2	1296.0	106.	103.0	104.5						
1926-27	747.2	854.7	1601.9	126.	131.8	129.2						
1927-28	524.0	572.0	1096.9	88.6	88.3	88.5						
1928-29	480.6	510.8	991.4	81.3	78.8	80.0	901.0	741.0	1642.0	92.5	105.8	98.3
1929-30	558.5	777.1	1335.6	94.5	119.8	107.7	1134.0	1242.0	2376.0	116.4	177.3	142.2
1930-31	927.9	878.9	1806.8	157.	135.5	145.7	1521.0	1108.0	2629.0	156.1	158.2	157.3
1931-32	391.5	643.1	1034.6	66.2	99.2	83.4	739.0	801.0	1540.0	75.8	114.4	92.2
1932-33	454.6	549.4	1004.0	76.9	84.7	81.0	846.0	707.0	1553.0	86.8	101.0	92.9
1933-34	486.6	501.0	987.6	82.3	77.3	79.7	739.0	474.0	1213.0	75.8	67.7	72.6
1934-35	552.2	566.9	1119.1	93.4	87.4	90.3	1047.0	862.0	1909.0	107.4	123.1	114.2
1935-36	732.1	716.8	1448.9	123.8	110.5	116.9	1338.0	922.0	2260.0	137.3	131.7	135.3
1936-37	612.1	542.7	1154.8	103.5	83.7	93.1	1196.0	687.0	1883.0	122.7	98.1	112.7
1937-38							760.0	554.0	1314.0	78.0	79.1	78.6
1938-39							943.0	685.0	1628.0	96.8	97.8	97.4
1939-40							1049.0	—	—	107.7	—	—
1940-41							1176.0	610.0	1786.0	120.7	87.1	106.9
1941-42							745.0	643.0	1388.0	76.5	91.8	83.1
1942-43							1174.0	672.0	1846.0	120.5	96.0	110.5
1943-44							701.0	561.0	1262.0	71.9	80.1	75.5
1944-45							1386.0	680.0	2066.0	142.2	97.1	123.6
1945-46							751.0	548.0	1299.0	77.1	78.3	77.7
1946-47							782.0	412.0	1194.0	80.2	58.8	71.5
1947-48							1069.0	722.0	1791.0	109.7	103.1	107.2
1948-49							466.0	373.0	839.0	47.8	53.3	50.2
1949-50												
Mean	591.3	648.5	1239.8				974.4	700.2	1670.9			

(mm and Percent of Normal)

	Lac d'Alfeld (alt. 620m)								
	mm			Percent of Normal			Average Percent of Normal		
Normal	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual
1925-26	1734.3	969.4	2703.7	120.5	100.1	111.5	113.2	101.6	108.0
1926-27	1511.2	1620.2	3131.4	105.0	168.5	129.1	115.5	150.2	129.2
1927-28	1055.7	781.5	1827.2	73.3	81.3	75.7	81.0	84.8	82.1
1928-29	1196.6	790.9	1987.5	83.1	82.3	81.9	85.6	89.0	86.7
1929-30	1525.1	1213.6	2738.7	106.0	126.2	112.9	105.6	141.4	120.9
1930-31	2366.1	1163.8	3529.9	164.4	121.0	145.5	159.2	138.2	149.5
1931-32	964.9	1043.7	2008.6	67.0	108.6	82.8	68.3	107.4	86.1
1932-33	1347.9	715.5	2063.4	93.6	74.4	85.1	85.8	120.0	86.3
1933-34	1114.3	596.5	1710.8	77.4	62.0	70.5	78.5	69.0	74.3
1934-35	1626.6	1147.1	2773.7	113.0	119.3	114.4	104.6	109.9	106.3
1935-36	1980.9	1180.8	3161.7	137.6	122.8	130.3	132.9	121.7	127.5
1936-37	1984.3	1045.7	3030.0	137.9	108.8	124.9	121.4	96.9	110.2
1937-38	1099.8	837.5	1937.3	76.4	87.1	79.9	77.2	83.1	79.2
1938-39	1406.6	1106.3	2512.9	97.7	115.1	103.6	97.2	106.4	100.5
1939-40	1830.3	1311.2	3141.5	127.2	136.4	129.5	117.4	136.4	129.5
1940-41	1970.2	966.1	2936.3	136.9	100.5	121.1	128.8	93.8	114.0
1941-42	1095.2	947.7	2042.9	76.1	98.6	84.2	76.3	95.2	83.6
1942-43	828.7	944.9	1773.6	57.6	98.3	73.1	89.0	97.2	91.8
1943-44	973.0	710.1	1683.1	67.6	73.9	98.4	69.8	77.0	72.4
1944-45	1967.9	835.3	2803.2	136.7	86.9	115.6	139.4	92.0	119.6
1945-46	1186.2	808.0	1994.2	82.4	84.0	82.2	79.8	81.2	80.0
1946-47	1163.3	601.1	1764.4	80.8	62.5	72.7	80.5	60.6	72.1
1947-48	1908.3	1206.0	3114.3	132.6	125.4	128.4	121.2	114.2	117.8
1948-49	705.9	529.5	1235.4	49.0	55.1	50.9	48.4	54.2	50.6
1949-50									
Mean	1439.3	961.4	2425.6						

TABLE 5 (*continued*)

Annual Precipitation on Lower Rhine

(mm and Percent of Normal)

	Johanniskreuz (alt. 472m)						Harmony of Main and Moselle Meurthe		
	mm			Percent of Normal			Percent of Normal		
Normal	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual	Oct.-Mar.	Apr.-Sept.	Annual
1925-26	476.0	414.0	890.0	105.5	91.2	97.8	— 7.7	—10.4	—10.2
1926-27	470.0	636.0	1106.0	104.1	140.1	121.5	—11.4	—10.1	— 7.7
1927-28	426.0	358.0	784.0	94.4	78.8	86.2	+13.4	— 6.0	+ 4.1
1928-29	372.0	306.0	678.0	82.4	67.4	74.5	— 3.2	—21.6	—12.2
1929-30	443.0	585.0	1028.0	98.2	128.9	113.0	— 7.4	—12.5	— 7.9
1930-31	691.0	615.0	1306.0	153.1	135.5	143.5	— 6.1	— 2.7	— 6.0
1931-32	287.0	575.0	862.0	63.6	126.7	94.7	— 4.7	+19.3	+ 8.6
1932-33	418.0	408.0	826.0	92.6	89.9	90.8	+ 6.8	—30.1	+ 4.5
1933-34	357.0	255.0	612.0	79.1	56.2	67.3	+ 0.6	—12.8	— 7.0
1934-35	442.0	504.0	946.0	97.9	111.0	104.0	— 6.7	+ 1.1	— 2.3
1935-36	470.0	551.0	1021.0	104.1	121.4	112.2	—28.8	— 0.3	—15.3
1936-37	518.0	483.0	1001.0	114.8	106.4	110.0	— 6.6	+ 9.5	— 0.2
1937-38	424.0	308.0	732.0	94.0	67.8	80.4	+16.8	—15.3	+ 1.2
1938-39	528.0	518.0	1046.0	117.0	114.1	115.0	+19.8	+ 7.7	+14.5
1939-40	631.0	585.0	1216.0	139.8	128.9	133.6	+20.4	— 7.5	+ 4.1
1940-41	621.0	553.0	1174.0	137.6	121.8	129.0	+ 8.8	+28.0	+15.0
1941-42	436.0	448.0	884.0	96.6	98.7	97.2	+20.3	+ 3.5	+13.6
1942-43	344.0	430.0	774.0	76.2	94.7	85.1	—12.8	— 2.5	— 6.7
1943-44	360.0	344.0	804.0	79.8	75.8	88.4	+10.0	— 1.2	+16.0
1944-45	—	—	—	—	—	—	—	—	—
1945-46	—	—	—	—	—	—	—	—	—
1946-47	314.0 Inc.	284.0	598.0 Inc.	69.6	62.6	65.7	—10.9	+ 2.0	— 6.4
1947-48	623.0	559.0	1182.0	138.0	123.1	129.9	+16.8	+ 8.9	+12.1
1948-49	278.0	270.0	548.0	61.6	59.5	60.2	+13.2	+ 5.3	+ 9.6
1949-50									
Mean	451.3	454.0	909.9					Percentage of Times within 15 percent	

TABLE 6

Precipitation and Runoff on Upper Rhine (above Basel)
Percent of Normal

Season	Snow Cover April 1 (Linth R. Basin)	Winter Precip. 6 Stations (Oct.-Mar.)	Winter Runoff at Basel (Oct.-Mar.)	Summer Runoff at Basel (Apr.-Sept.)	Summer Precip. 6 Stations (Apr.-Sept.)	Divergence of Summer Runoff (a) From Snow Cover Apr..)	(b) From Winter Precip.)	(c) From Summer Precip.
1935-36		143.4	149.8	112.8	122.2		—30.6	— 9.4
1936-37		114.3	130.3	121.2	98.9		+ 6.9	+22.3
1937-38		82.3	99.2	90.2	95.9		+ 7.9	— 5.7
1938-39		79.2	84.9	119.6	107.4		+40.4	+12.2
1939-40		135.6	162.5	114.4	116.7		—21.2	— 2.3
1940-41		96.0	135.1	98.5	92.1		+ 2.5	+ 6.4
1941-42		76.5	97.8	88.4	87.7		+11.9	+ 0.7
1942-43	72.0	75.4	87.5	80.6	93.3	+ 8.6	+ 5.2	—12.7
1943-44	142.0	75.8	73.8	93.5	82.0	—48.5	+17.7	+11.5
1944-45	118.1 (Precip. 169.1)	145.0	156.6	98.7	96.9	—19.4	—46.3	+ 1.8
1945-46	102.4	93.8	106.6	100.3	100.0	— 2.1	+ 6.5	+ 0.3
1946-47	72.4	81.0	87.8	72.3	69.5	— 0.1	— 8.7	+ 2.8
1947-48	93.1 (Precip. 152.1)	127.1	112.4	95.4	98.4	+ 2.3	—31.7	— 3.0
1948-49	74.5			69.8	(Apr.-Sept.)	— 4.7		
1949-50	92.3			77.5		—14.8		
							Harmony Within 15 Percent	
						75	54	90

The snow cover was found to harmonize with the summer runoff within 15 percent on 75 percent of the eight seasons available but the winter precipitation maintained the same harmony on only 54 percent of thirteen seasons.

As expected, the winter precipitation and the winter runoff harmonized within 15 percent only 40 percent of the time. On the other hand, the summer precipitation and summer runoff harmonized 9 times out of 10.

Since the runoff was measured only at Basel, the lower lands of Switzerland and the large lakes Constance, Zürich, Lucerne and Neuchatel entered into the forecast.

In Table 7 by Dr. E. Hoeck on Runoff in the Basins of the Upper Rhine (April 1 — July 31), the comparison for 1927-1950 is based largely on precipitation but is confined to the Rhine Source, the Linth, Reuss and Aar mostly above large lakes and the period of summer runoff is reduced to the four months of April-July. Weighting factors also are used. The total Swiss Rhine at Basel or Rheinfelden concludes the series.

On the same basis of 15 percent allowable divergence, harmony within this limit between forecasted and measured flow was attained in 90 percent of the years in each of the four basins. However, for the total watershed, harmony within 15 percent was attained in only 62 percent of the forecasts. The latter result is comparable to that in Table 6 but based on the six months period of flow.

The extension of snow surveys to the lower Swiss valleys and weighting the expected contribution of all feeders to the main Rhine, as in the forecast of the Colorado River on the American Continental Divide, may produce a forecast as accurate as those of the four sub-basins in the table.

Forecast Problems on the Lower Rhine

The prevalence of rain on the Lower Rhine and the inefficiency of rain measurements as a stable index of runoff practically nullifies even approximate forecasts in that area. The runoff efficiency of 29.7 percent of precipitation on the Main and 24.2 percent on the Moselle-Meurthe indicates the severe wastage enroute.

This is emphasized by the following list of streams of varying efficiency in which the Rhine is included.

Runoff Efficiency

Character of Precipitation	Stream	Percent Efficiency (Annual)
HeavySnow 48.2 in	Upper Rhine	72.9
» » 60 in.	South Yuba (U.S.A.)	78.2
Heavy Rain-Snow 50-75 in.	Rogue »	51.8
Medium Snow-Rain	Merrimac »	48.3
» » » 40-50 in.	Susquehanna »	53.6
Mostly Rain35.8 in.	Main (Germany)	29.7
69.2 in.	Moselle-Meurthe (France)	24.2
30-40 in.	Wisconsin (U.S.A.)	51.0
20-40 in.	Upper Mississippi (U.S.A.)	23.7
20-30 in.	Red River of the North (U.S.A.)	5.7

(Percent of Normal)

1927-1951. Forecasted and Measured by Dr. E. Hoeck

Année	Rhin St. Margrethen			Linth Weesen-Biäsche			Reuss Lucerne			Aar Berne			Rhin Rheinfelden		
	Prév. %	Mesure %	Diff. %	Prév. %	Mesure %	Diff. %	Prév. %	Mesure %	Diff. %	Prév. %	Mesure %	Diff. %	Prév. %	Mesure %	Diff. %
1927	120.5	124.3	+ 3.8	121.7	133.5	+11.8	123.5	128.6	+ 5.1	122.0	118.0	— 4.0	121.6	123.7	+ 2.1
1928	88.1	94.9	+ 6.8	78.7	84.5	+ 5.8	87.7	93.5	+ 5.8	97.5	98.3	+ 0.8	93.7	85.6	— 8.1
1929	103.5	85.8	*—17.7*	84.8	93.3	+ 8.5	100.6	99.6	— 1.0	102.1	103.8	+ 1.7	102.7	86.7	—16.0
1930	91.9	88.1	— 3.8	83.3	90.5	+ 7.2	97.4	101.6	+ 4.2	94.0	117.1	*+23.1*	93.1	111.0	+17.9
1931	114.6	115.1	+ 0.5	117.7	117.5	— 0.2	118.3	117.6	— 0.7	121.8	119.7	— 2.1	119.2	120.2	+ 0.8
1932	85.2	90.6	+ 5.4	79.1	84.5	+ 5.4	86.8	93.2	+ 6.4	88.4	97.1	+ 8.7	87.0	103.6	+16.6
1933	83.4	93.4	+10.0	85.1	101.3	+16.2	84.3	98.6	+14.3	90.9	86.5	— 4.4	87.8	97.1	+ 9.3
1934	86.5	71.0	—15.5	78.6	60.7	—17.9	90.6	75.4	—15.2	91.1	74.0	—17.1	89.1	64.1	*—25.0*
1935	126.6	122.9	— 3.7	127.3	119.5	— 7.8	121.7	116.5	— 5.2	110.8	110.5	— 0.3	117.2	118.1	+ 0.9
1936	115.5	112.5	— 3.0	110.0	102.5	— 7.5	116.2	111.5	— 4.7	105.5	110.4	+ 4.9	109.6	112.8	+ 3.2
1937	109.9	115.6	+ 5.7	118.5	123.2	+ 4.7	106.2	123.2	*+17.0*	111.2	118.1	+ 6.9	110.9	121.2	+10.3
1938	95.7	89.4	— 6.3	95.0	98.1	+ 3.1	93.9	93.0	— 0.9	89.1	89.2	+ 0.1	91.6	89.9	— 1.7
1939	94.9	108.9	+14.0	98.3	99.3	+ 1.0	96.1	103.8	+ 7.7	96.5	105.2	+ 8.7	95.8	117.8	+22.0
1940	106.5	115.5	+ 9.0	120.8	123.0	+ 2.2	107.2	109.9	+ 2.7	106.3	106.7	+ 0.4	106.5	114.6	+ 8.1
1941	95.8	93.6	— 2.2	94.3	84.4	— 9.9	90.9	86.9	— 4.0	96.0	94.4	— 1.6	95.9	98.0	+ 2.1
1942	84.8	81.6	— 3.2	97.5	86.5	—11.0	92.5	81.8	—10.7	88.9	89.8	+ 0,9	87.1	88.4	+ 1.3
1943	85.2	88.0	+ 2.8	93.6	94.5	+ 0.9	82.8	80.2	— 2.6	86.9	81.9	— 5.0	86.1	80.9	— 5.2
1944	97.0	99.8	+ 2.8	114.1	111.0	— 3.1	106.0	101.2	— 4.8	97.8	90.3	— 7.5	97.5	94.0	— 3.5
1945	116.6	118.0	+ 1.4	127.4	121.0	— 6.4	113.7	111.1	— 2.6	113.4	110.1	— 3.3	114.8	99.5	—15.3
1946	107.6	110.0	+ 2.4	99.9	96.4	— 3.5	94.3	99.1	+ 4.8	96.2	97.1	+ 0.9	100.7	100.1	— 0.6
1947	90.2	81.0	— 9.2	93.8	74.8	*—19.0*	90.0	73.7	—16.3	93.6	81.8	—11.7	92.1	72.9	—19.2
1948	109.9	107.6	— 2.3	104.3	103.6	— 0.7	96.5	87.8	— 8.7	92.0	83.6	— 8.4	99.1	95.1	— 4.0
1949	86.7	69.4	—17.3	87.1	75.0	—12.1	84.6	73.0	—14.6	88.8	73.4	—15.4	87.8	69.8	—18.0
1950	100.2	87.7	—12.5	93.5	79.0	—14.5	93.9	87.2	— 6.7	91.6	86.9	— 4.7	94.9	77.5	—17.4
1951	138.5			129.5			125.9			116.7			125.6		
Harmony within 15%			90			90			90			90			62
Percent of Total															

TABLE 8

Precipitation and Runoff on Main River (above Schweinfurt)

	(Percent of Normal)						
	Winter Precip. (Oct.-Mar.)	Winter Flow (Oct.-Mar.)	Summer Flow (Apr.-Sept.)	Summer Precip. (Apr.-Sept.)	(a) Winter Prec. from Winter Flow	Divergence (b) Winter Precip. from Summer Flow	(c) Summer Precip. from Summer Flow
1924-25		97.8	90.7				
1925-26	105.5	137.8	146.5	91.2	+32.2	+41.0	+55.3
1926-27	104.1	93.3	165.1	140.1	—10.8	+61.0	+25.0
1927-28	94.4	127.4	82.6	78.8	+33.0	—11.8	+ 3.8
1928-29	82.4	63.7	62.8	67.4	—18.7	—19.6	— 4.6
1929-30	98.2	48.9	93.0	128.9	—49.3	— 5.2	—35.9
1930-31	153.1	155.6	118.6	135.5	+ 2.5	—34.5	—16.9
1931-32	63.6	81.5	94.2	126.7	+17.9	+30.6	—32.5
1932-33	92.6	78.5	83.7	89.9	—14.1	— 8.9	— 6.2
1933-34	79.1	44.4	36.0	56.2	—34.7	—43.1	—20,2
1934-35	97.9	80.0	103.5	111.0	—17.9	+ 5.6	— 7.5
1935-36	104.1	88.9	123.2	121.4	—15.2	+19.1	+ 1.8
1936-37	114.8	157.0	109.3	106.4	+42.2	— 5.5	+ 2.9
1937-38	94.0	70.4	57.0	67.8	—23.6	—37.0	—10.8
1938-39	117.0	87.4	74.4	114.1	—29.6	—42.6	—39.7
1939-40	139.8	145.2	148.8	128.9	+ 5.4	+ 9.0	+19.9
1940-41	137.6	143.0	180.2	121.8	+ 5.4	+42.6	+58.4
1941-42	96.6	137.0	95.3	98.7	+40.4	— 1.3	— 3.4
1942-43	76.2	63.0	60.5	94.7	—13.2	—15.7	—34.2
1943-44	79.8	68.9	112.8	75.8	—10.9	+33.0	+37.0
1944-45	—	130.4	109.3	—	—	—	—
1945-46	—	100,0	74.4	—	—	—	—
1946-47	64.6	78.5	60.5	62.6	+ 8.9	— 9.1	— 2.1
1947-48	138.0	—	—	123.1			
1948-49	61.6			59.5			

Harmony Within 15 Percent (Pct. of Total)

TABLE 9

Precipitation and Runoff on Moselle-Meurthe Rivers (above Kochem)

(Percent of Normal)

	Winter Precip. Oct.-Mar. (—0.17%)	Winter Flow Oct.-Mar.	Summer Flow Apr.-Sept.	Summer Precip. Apr.-Sept. (—5.5%)	Weighted *	Divergence (a) Winter Precip. from Winter flow	Divergence (b) Winter Precip. from Summer flow	Divergence (c) Summer Precip. from Summer flow
1936-37	121.4	133.9	80.1	96.9	102.5	+12.5	— 41.3	—22.4
1937-38	77.2	73.6	27.0	83.1	87.9	— 3.6	— 50.2	—60.9
1938-39	97.2	83.3	62.3	106.4	112.5	—13.9	— 34.9	—50.2
1939-40	117.4	151.7	80.4	136.4	144.3	+34.3	— 37.0	—63.9
1940-41	128.8	110.4	78.9	93.8	99.2	—18.4	— 49.9	—20.3
1941-42	76.3	80.0	123.5	95.2	100.7	+ 3.7	+ 47.2	+22.8
1942-43	89.0	64.5	138.2	97.2	102.8	—24.5	+ 49.2	+35.4
1943-44	69.8	86.8	176.5	77.0	81.5	+17.0	+106.7	+95.0
1944-45	139.4	108.9	88.0	92.0	97.4	—30.5	— 51.4	— 9.4
1945-46	79.8	86.4	78.9	81.2	85.9	+ 6.6	— 0.9	— 7.0
1946-47	80.5	125.4	138.6	60.6	64.1	+44.9	+ 58.1	+74.5
1947-48	121.2	94.6	121.7	114.2	120.8	—26.6	— 0.5	+ 0.9
						Harmony Within 15 Percent (Pct. of Total)		
*Normal Raised 5.5% on basis of data 1925 to 1949.						42	17	25

The Main

To avoid the effect of diversions, measurements have been confined to the upper part of the stream. But even so, divergence is found to exist (1) between the winter precipitation and the winter flow, (2) between the winter precipitation and the summer flow, and (3) even between the summer precipitation and the summer flow. In the first two comparisons harmony within 15 percent occurs only 40 percent of the time and in the third only 45 percent of the time (Table 8).

The Moselle Meurthe

In two of the three similar comparisons on the Moselle-Meurthe the divergence is even greater, the harmony within 15 percent being only 42, 17 and 25 percent respectively (Table 9) as compared with 40, 40 and 45 for the Main .

TABLE 10

Comparison of Summer Runoff (Apr.-Sept.) of Upper Rhine (Basel), Main, and Moselle-Meurthe Rivers

(Percentage of Normal)

Years	Rhine (Basel) (1808-1947) 1290 m^3/s	Main (Schweinfurt) (1935-1947) 86 m^3/s	Moselle-Meurthe (Kochem) (1937-1948) 332 m^3/s	Total Rhine (1936-1940) (Rees) 2352 m^3/s
1924-25		90.7		
1925-26		146.5		
1926-27		165.1		
1927-28		82.6		
1928-29		62.8		
1929-30		93.0		
1930-31		118.6		
1931-32		94.2		
1932-33		83.7		
1933-34		36.0		
1934-35		103.5		
1935-36	112.8	123.2		
1936-37	121.2	109.3	80.1	
1937-38	90.2	57.0	27.0	
1938-39	119.6	74.4	62.3	
1939-40	114.4	148.8	80.4	
1940-41	98.5	180.2	78.9	
1941-42	88.4	95.3	123.5	
1942-43	80.6	60.5	138.2	
1943-44	93.5	112.8	176.5	
1944-45	98.7	109.3	88.0	
1945-46	100.3	74.4	78.9	
1946-47	72.3	60.5	138.6	
1947-48	95.4		121.7	

Accuracy of forecasts of summer flow within 15 percent for the Lower Rhine is therefore narrowed to 40 percent of the years on the Main and to only 17 percent on the Moselle-Meurthe.

Comparison of Summer Runoff Percentages on the Upper Rhine, Main and Moselle-Meurthe

The seasonal trends of the three main feeders of the Rhine rarely run closely parallel (Table 10). Yet during a 12-year comparison of the output of the Swiss feeders at Basel and the German-French feeders at Rees (Table 11), the average divergence between seasonal percentages has been only 27 percent and the maximum 51 percent. The volumes of the two (upper and lower) sources are closely similar, being 1290 and 1464 m^3/s (Total 2352 m^3/s). The maximum probable correction (either + or —) to apply to the forecast of joint flow at Rees would therefore be 51 percent of 1464 m^3/s or 747 m^3/s. It could well be less.

The forecast base must, therefore, be the summer flow of Swiss water modified by the German-French flow downstream. The determination of the former should be relatively precise but trends of the latter should be visible.

TABLE 11

Comparison of Summer Runoff (Apr.-Sept.) in Upper and Lower Rhine

	(*Percentage of Normal*)		
Years	1. Upper Rhine Rhine at (Base[) 1290 m^3/s (Measured)	2. Lower Rhine Main, Moselle Meurthe and Other Sources (Rees) 1464 m^3/s (Estimated from Main and Moselle-Meurthe) Unweighted average	Divergence from Upper Rhine at Basel (Pct. of 1464 $m^3/$)
1936-37	121.2	94.7	— 26.5
1937-38	90.2	42.0	— 48.2
1938-39	119.6	68.4	— 51.2
1939-40	114.4	114.6	+ 0.2
1940-41	98.5	129.6	+ 31.1
1941-42	88.4	109.4	+ 21.0
1942-43	80.6	99.4	+ 18.8
1943-44	93.5	144.6	+ 51.1
1944-45	98.7	98.6	— 0.1
1945-46	100.3	76.6	— 23.7
1946-47	72.3	99.6	+ 27.3
1947-48	95.4	—	
	Base flow	Supplementary Flow	

The Purpose of Forecasts

In the mountain area of Europe, the generation of power and the forecasting, of firm streamflow should be a major activity. Irrigation, on the other hand, because of heavy and well-distributed precipitation (Table 1) should be a minor interest except in specialized agriculture or local need of fresh water as in the Netherlands.

Floods and navigation would be an emergency interest, served better by a study of climate and weather than by the usual stream-level gages.

Ijssel Meer (Ijssel Lake)*

Ijssel Lake, in the Rhine delta, is the only storage of fresh water in Holland, There is plenty of water but in the lower part of the country it is mostly brackish. Holland, therefore, is as badly in want of fresh water as many semi-arid countries.

It was originally the southern lobe of the Zuider Zee and was reclaimed from the northern portion by an enclosure dam. It covered 3,700 Km^2. The shallower rim of the lake is being progressively dammed off into agricultural «polders» until ultimately its final area will be about 1,300 Km^2. But it will water about 6,500 Km^2 (approximately 1/2m deep).

In summer the level of the water in the lake is kept at 0.15 m (6 in.) below mean sea level; in winter it is lowered 0.25 m to 0.40 m (16 in.) below m.s.l. The disk (325,000,000 M^3) between the two is available for supply of the land around the lake. The surplus of water is made to flow into the sea through evacuation sluices in the enclosure dam. They are operated at low tide: the sea level is then 0.9 m (36 in.) below m.s.l.

When all the polders have been reclaimed, the mean depth of the lake will be about 4 meters below m.s.l. and the water content about 5,000,000,000 cubic meters (4,052,000 acre feet).

The turn-over is slightly more than once a year; it will increase to 2 ½ times.

The quantity of water needed annually for agriculture is estimated at present as 3,600,000,000 m^3, one-third of this being required during the winter 6 months of October-March. The peak use is 1,000,000,000 during the two summer months of June-July.

It may increase in future. In dry summers the supply is less than the demand. In those cases the storage must be drawn upon: the level of the lake is lowered as the summer proceeds. Now about 700,000,000 m^3 may be got in this way. In future, when the lake is smaller the level in summer will be increased: the amount to be stored will remain the same.

Only one branch of the Rhine delta (The Ijssel, from which the lake has received its name) flows into Ijssel Lake. It is the minor one, carrying only 10-15 percent (239-359 m^3/s) of the discharge (Table 1).

The mean value of the runoff of the Ijssel is about 250 M 3/Sec. Slightly more (300) in January-March (German water), about normal in May-August (Swiss water), lowest (200) in September-November (no more melting, not yet much rain). A very high value (one day per year) is 1200 m^3/s; a low one (also one day per year) is 75 m $^3/s$. Highest and lowest observed are 2200 and 40.

The runoff of the Ijssel River is the main source of supply, but there are others: small rivers and brooks bringing water from higher grounds, pumping stations and sluices bringing water from the «polders». The rain on the lake itself is also important but on the other hand the evaporation may not be neglected.

For a «normal» year:

Ijssel River	8,500,000,000 m^3	
Higher grounds	3,400,000,000	
Polders	1,500.000,000	
Rain	2,000,000,000	(26.24 in.)
Deduct evaporation	— 2,000,000,000	»
Total . .	13,400,000,000	

*From a letter by Dr. Thijsse, 13 Nov., 1951.

One of the difficult items is how to choose the right moment for raising the level of the lake in spring by interrupting the evacuation. If it is done late and the summer is dry, it may not be possible to collect the water needed for storage: in that case the summer level will not be reached. If, on the other hand, the start is made too early, the surrounding land may get water-logged because of interference with the evacuation.

Therefore, we are greatly interested in Dr. Hoeck's predictions. We want to know about March 20 if there is so much snow in the catchment basin that a very low runoff in the next summer will be impossible. If this is the case, we may wait longer before raising the level of the lake.

For the German tributaries a long-term prediction is impossible, because the lapse of time between the moment a drop of rain falls and the moment when its influence is felt in Holland is too short (one or two weeks). Here we use the daily water-levels along the Rhine to predict a flood. If we know a flood will come, we may anticipate it and open the sluices in order to make room for it in the lake. I told about this in 1948 at Oslo.

In the near future weirs will be constructed in the middle branch of the Rhine (Nederrijn) in order to divert a higher percentage of the total flow to the Ijssel (and Ijssel Lake) in periods of low runoff.

The total annual normal flow of the Rhine at Rhees is 2390 m^3/s or 75,371,040,000 m^3, or more than 5 times the total normal supply from all sources for the lake. The total annual normal flow of the Ijssel is 250 m^3/s or an estimated 8,500,000,000 m^3. The channel can carry 1200 m^3/s or 37,843,200,000 m^3, or ½ the normal supply of the Rhine. The weirs and the water rights would be within the domain of Holland. Control could conceivably be exercised at the weirs to divert scarce water into the Ijssel or flood waters back into the main Rhine. In other words, the control of Ijssel Lake would begin at the weirs.

In 1946-47 (Table 10) a drought* was experienced at Ijssel Lake with a summer (April-September) runoff in the Swiss Rhine (Basel) of 72.3 percent of normal and 60.5 and 138.6 percent respectively in the Main and Moselle-Meurthe. The Swiss Rhine alone flowed 933 m^3/s or 14,711,544,000 m^3 for the 6 months period and the German-French feeders were approximately normal. The weirs have, therefore, become an emergency project.

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FORECASTING THE SUMMER RUNOFF OF THE RHINE RIVER

J. E. CHURCH

THE NECKAR BASIN*

The River Neckar should be included in any study of the Lower Rhine Basin as a source of further information as well as a source of water. It is a close companion of the Main but benefits more from highland snow.

Hydrology

Both rise on the flank of the Danube which circles north of Switzerland, but the Main lies farther to the east and north.

The area of the Neckar watershed is 14,000 Km^2, the Main, 27,395 Km^2, and the Moselle-Meurthe, 28,156 Km^2. But owing to diversion by canals only the upper 4,002 Km^2 (29%) of the Neckar and 12,719 Km^2 (46%) of the Main have been gaged. The runoff of their entire basins, however, would seem to be in proportion to the area. The unmeasured area (24,674 Km^2) is 15.5 percent of the entire Rhine Basin to Rees. This reduces other downstream sources of runoff to 58,209 Km^2 or 36.5 percent of the total. Of this, only 1,056 Km^2 or less than one percent belongs to the Moselle-Meurthe. Of these three feeders, the Main and the Moselle-Meurthe are almost identical in area and the Neckar is half the size of each.

A map is highly desirable but the revised «Hydrology of the Rhine Basin» is included as Table 1.

Precipitation

The mean seasonal precipitation in the Neckar Basin is represented by three stations of Rottweil, Tübingen, and Heilbronn, located in above order from source toward outlet. Their trends are reasonably similar and the mean seasonal percentage fits into the pattern of decreasing winter and increasing summer precipitation toward Switzerland. See Table 2.

Efficiency

Efficiency is the relation of runoff (i.e. discharge) to precipitation on the watershed and is increased by the presence of a seasonal snow cover. Tables 3*a* and 3*b* are illustrative.

There are all too few precipitation stations for a close estimate but the trends are obvious. The snow-fed streams are far more efficient. The Neckar appears to hold its snow fairly well in winter but has lost it by summer.

Snow is doubly efficient over rain in production of net or total runoff because

(*) Material for this study has kindly been furnished by Dr. W. Friedrich of the Bundesanstalt für Gewässerkunde at Koblenz, Germany.

TABLE 1

The Hydrology of the Rhine Basin

Divisions	Area Km²	Percent of Total	Elevation of Stations (m. above sea)	Mean Precipitation winter (mm.)	(Oct.-Mar.) (% of Ann.)	Summer (mm.)	Apr.(-Sept.) (% of Ann.)	Annual (mm.)
1. Upstream								
Swiss Rhine (Basel)	35,925	22.5	453 — 2323	468	37.6	767	62.4	1236
2. Downstream								
Neckar	4,002*	2.5		285	39.0	445	61.0	730
Main (Schweinfurt)	12,719	8.0	472	457	49.8	461	49.8	918
Moselle-Meurthe (Kochem)	27,100	17.0	77.5 370 — 620	993	54.7	782	45.3	1775
Other Sources	79,937	50.0		725	53.9	622	46.2	1346
	123,758							
3. Entire Rhine (Rees)	159,683	100.0		639	48.8	670	51.1	1310

*Measured area 4,002 Km² (=29%) Total 14,000 Km²

TABLE 1 (*continued*)

The Hydrology of the Rhine Basin

Divisions	Mean Runoff (In.)	Winter (m³/s)	(Oct.-Mar.) (% of Ann.)	Summer (m³/s)	(Apr.-Sept.) (% of Ann.)	Annual (m³/s)	(% of Entire Stream)		Efficiency of Runoff Annual Only (% of Precip.)	
1. Upstream										
Swiss Rhine (Basel)	48.2	763.0	37.2	1290.0	62.8	1027.0	43.0		72.9	Oct.-Mar. 67.7% Apr.-Sept. 67.7%
2. Downstream										
Neckar	28.7	47.4	55.3	38.3	44.7	42.8*	1.8	57.0	46.2	Oct.-Mar 65.3%
Main(Schweinfurt)	35.8	135.0	61.3	86.0	38.8	110.0	4.6		29.7	Apr.-Sept. 33.9%
Moselle-Meurthe (Kochem)	69.2	425.0	56.1	332.0	44.0	384.0	16.1		24.2	25.8
Other Sources	52.5	1307.0	66.8	644.0	33.0	869.0	34.5		24.6	
3. Entire Rhine (Rees)	51.1	2630.0	52.8	2352.0	47.2	2390.0 (= 61,064,500 AF)	100.0		36.0	

*For entire area of 14,000 Km², estimated runoff is 148 M 3/S = 6.2% of total flow of Rhine at Rees.
For the total Main, the estimate should be 237 M 3/S = 9.9%.
For «other sources» 594 M 3/S = 24.8%.

Thus, Swiss Rhine 43.0%
Neckar 6.2%
Main 9.9%
Moselle-Meurthe 16.1%
Other Sources 24.8%

fewer primings of the soil are required. This seems to be illustrated in the following comparison of the Neckar and the Moselle-Meurthe.

TABLE 2

Mean Seasonal Precipitation Trends

	Quantitative MM			Percentage of Annual		
	Winter Oct.-Mar.	Summer Apr.-Sept.	Ann.	Winter Oct.-Mar.	Summer Apr.-Sept.	Ann.
(a) Neckar Basin						
Rottweil	312	468	780	40.0	60.0	100
Tübingen	252	460	712	35.4	64.6	100
Heilbronn	292	406	698	41.8	58.2	100
Average	285	445	730 = 28.7 in.	39.1	60.9	100
(b) Rhine Basins						
*Swiss Rhine	468	767	1236	37.6	62.4	100
*Neckar	285	445	730	39.1	60.9	100
*Main	457	461	918	49.8	49.8	100
*Moselle-Meurthe	993	782	1775	54.7	45.3	100

*Read upward.

TABLE 3*a*

General Stream Efficiency
(Percent of Streamflow to Precipitation)

	Annual	
Swiss Rhine	72.9	Percent
Neckar	46.2	»
Main	29.7	»
Moselle-Meurthe	24.2	»
Entire Rhine	36.0	»

TABLE 3*b*

Seasonal Stream Efficiency Neckar, Moselle-Meurthe, Swiss Rhine
(Percent of Annual)

	Winter: Oct.-Mar.	Summer: Apr.-Sept.
1. Neckar (Snow dominates in winter but gone early).		
Precipitation	39.0	61.0
Runoff	55.3	44.7
Efficiency	65.3	33.9
2. Moselle-Meurthe (Only occasional snow)		
Precipitation	54.7	45.3
Runoff	56.1	44.0
Efficiency	24.2	25.8
3. Swiss Rhine (Snow through winter and spring).		
Precipitation	37.6	62.4
Runoff	37.2	62.8
Efficiency	67.7	67.7

The runoff of the Swiss Rhine in winter seems large. Can the lowlands of Switzerland and its lakes account for this?

Harmony of Precipitation

During the quarter century of record, fair harmony of precipitation both seasonal and annual has been noted between the Neckar and the Main and the Main and the Moselle-Meurthe — a bit closer even between the latter than the former although situated farther apart.

For example, the percentage of times when harmony has been within 15 per cent is:

	Oct.-Mar.	Apr.-Sept.	Annual
Neckar and Main	86	64	82
Main and Moselle-Meurthe	73	77	91

Details by years for the Main and Moselle-Meurthe are given in Table 5 of the original paper. Those for the Neckar and Main are in the following Table 4.

Year	Oct.-Mar.		Apr.-Sept.		Annual		Harmony between Neckar and Main		
	Neckar	Main	Neckar	Main	Neckar	Main	Oct.-Mar.	Apr.-Sept.	Annual
1920-21	46.7		68.6		60.0				
1921-22	93.2		128.9		114.8				
1922-23	117.9		67.6		87.0				
1923-24	115.3		126.5		112.2				
1924-25	65.5		94.6		88.8				
1925-26	101.8	105.5	95.1	91.2	98.1	97.8	3.7	3.9	0.3
1926-27	101.6	104.1	121.9	140.1	139.3	121.5	2.5	18.2	17.8
1927-28	85.1	94.4	65.7	78.8	73.4	86.2	9.3	13.1	12.8
1928-29	92.8	82.4	82.0	67.4	86.4	74.5	10.4	14.6	11.9
1929-30	91.7	98.2	127.1	128.9	113.2	113.0	6.5	1.8	0.2
1930-31	143.9	153.1	154.0	135.5	150.0	143.5	9.2	18.5	6.5
1931-32	67.4	63.6	107.8	126.7	82.1	94.7	3.8	18.9	12.6
1932-33	74.3	92.6	90.6	89.9	84.0	90.8	18.3	0.7	6.8
1933-34	68.5	79.1	98.9	56.2	87.3	67.3	10.6	42.7	20.0
1934-35	100.5	97.9	92.9	111.0	96.0	104.0	2.6	18.1	8.0
1935-36	127.4	104.1	122.4	121.4	124.4	112.2	23.3	1.0	12.2
1936-37	128.7	114.8	92.6	106.4	106.8	110.0	13.9	13.8	3.2
1937-38	93.3	94.0	118.8	67.8	110.2	80.4	0.7	51.0	29.8
1938-39	109.4	117.0	150.3	114.1	134.6	115.0	7.6	36.2	19.6
1939-40	159.3	139.8	139.7	128.9	147.5	133.6	19.5	10.8	13.9
1940-41	123.1	137.6	109.0	121.8	116.5	129.0	14.3	12.8	12.5
1941-42	110.9	96.6	95.7	98.7	101.5	97.2	14.3	3.0	4.3
1942-43	70.9	76.2	89.6	94.7	82.2	85.1	5.3	5.1	2.9
1943-44	78.1Inc	79.8	80.4	75.8	79.6	88.4	1.7Inc	4.6	8.8
1944-45	138.7 »	—	123.2Inc	—	129.7Inc	—	—	—	—
1945-46	101.0 »	—	101.0 »	—	101.0 »	—	—	—	—
1946-47	76.4 »	69.6	65.2 »	62.6	69.3 »	65.7	6.8Inc	2.6Inc	3.6Inc
1947-48	147.9	138.0	98.1	123.1	117.7	129.9	9.9	25.0	12.2
1948-49	63.2	61.6	71.6	59.5	68.3	60.2	1.6	12.1	8.1
1949-50	86.4		112.3		102.0		86.0	64.0	82.0

Percentage of times within 15 per cent.

TABLE 5

Variation Between Precipitation and Runoff in Neckar Basin Percentage of Precipitation and Runoff (Based on Normal)

Year	Rottweil			Tübingen			Heilbronn		
	Oct.-Mar.	Apr.-Sept.	Ann.	Oct.-Mar.	Apr.-Sept.	Ann.	Oct.-Mar.	Apr.-Sept.	Ann.
1920-21	37.2	69.4	56.5	44.8	60.4	54.9	58.2	76.1	68.6
1921-22	101.9	140.0	124.7	94.4	125.4	114.5	83.2	121.2	105.3
1922-23	114.7	68.6	87.1	126.2	64.3	86.2	112.7	70.0	87.8
1923-24	133.0	123.3	127.2	111.1	124.3	119.7	102.4	132.0	119.6
1924-25	60.6	99.6	84.0	57.9	92.2	81.2	78.1	92.1	86.2
1925-26	111.9	106.0	108.3	92.1	105.0	100.4	101.4	74.4	85.7
1926-27	106.1	125.6	117.8	100.8	119.3	112.8	97.9	120.7	111.2
1927-28	76.3	61.8	67.6	86.1	68.7	74.9	92.8	66.7	77.7
1928-29	90.4	83.5	86.3	95.2	90.2	92.0	92.8	72.2	80.8
1929-30	92.0	110.0	102.8	102.0	138.5	125.6	81.2	132.8	111.2
1930-31	144.6	126.7	133.8	157.1	172.2	166.9	130.1	163.1	149.3
1931-32	67.6	138.8	82.6	75.8	75.2	75.4	58.9	109.4	88.3
1932-33	78.2	92.9	87.1	70.6	76.5	74.4	74.0	102.5	90.5
1933-34	76.3	110.9	97.1	65.5	103.9	90.3	63.7	82.0	74.4
1934-35	102.9	87.4	93.6	97.2	96.1	96.5	101.4	95.3	97.9
1935-36	135.9	128.4	131.4	125.4	125.2	125.3	120.9	113.5	116.6
1936-37	134.9	98.5	113.1	121.4	86.5	98.9	129.8	92.9	108.3
1937-38	89.7	134.2	116.4	96.8	118.9	111.0	93.5	103.4	103.3
1938-39	92.0	162.0	134.0	114.9	161.5	144.9	121.2	127.3	124.8
1939-40	151.3	134.8	141.4	157,1	143.3	148.2	169.5	140.9	152.9
1940-41	126.3	100.4	110.8	125.4	119.1	121.3	117.5	107.6	117.5
1941-42	94.9	93.2	93.8	114.3	88.9	97.9	123.6	104.9	112.8
1942-43	76.6	85.9	82.2	66.3	83.7	77.5	69.9	99.3	87.0
1943-44	80.4	71.4	75.0	71.8Inc	80.0	77.1	82.2	89.9	86.7
1944-45	—	—	—	—	—	—	138.7	123.2	129.7
1945-46	—	—	—	—	—	—	101.0	101.0	101.0
1946-47	—	—	—	82.5	67.2	72.6	70.2	63.1	66.0
1947-48	112.5	88.7	98.2	150.8	97.6	116.4	180.5	108.1	138.4

Year	Average			Neckar River			Departure of Runoff from Precipitation, Based on Average of Rottweil, Tubingen Heilbronn		
	Oct.-Mar.	Apr.-Sept.	Ann.	Oct.-Mar.	Apr.-Sept.	Ann.	Oct.-Mar.	Apr.-Sept.	Ann.
1920-21	46.7	68.6	60.0	46.4	28.7	37.4	— 0.3	—39.9	—22.6
1921-22	93.2	128.9	114.8	63.3	161.9	107.5	—29.9	+ 33.0	— 7.3
1922-23	117.9	67.6	87.0	181.4	67.9	130.8	+63.5	+ 0.3	+43.8
1923-24	115.3	126.5	112.2	99.2	182.8	135.5	+ 16.1	+ 56.3	+ 23.3
1924-25	65.5	94.6	83.8	75.9	86.2	79.4	+ 10.4	— 8.4	— 4.4
1925-26	101.8	95.1	98.1	109.7	120.1	114.5	+ 7.9	+25.0	+16.4
1926-27	101.6	121.9	139.3	75.9	134.1	102.8	—25.7	+12.2	—36.5
1927-28	85.1	65.7	73.4	97.0	44.4	74.8	+11.9	—21.3	+ 1.4
1928-29	92.8	82.0	86.4	50.6	54.8	51.4	—42.2	—27.2	—35.0
1929-30	91.7	127.1	113.2	42.2	94.0	65.4	—49.5	—33.1	—47.8
1930-31	143.9	154.0	150.0	145.6	177.5	158.9	+ 1.7	+23.5	+ 8.9
1931-32	67.4	107.8	82.1	86.3	101.8	93.5	+19.1	— 6.0	+11.4
1932-33	74.3	90.6	84.0	52.7	80.9	65.4	—21.6	— 9.7	—18.6
1933-34	88.5	98.9	87.3	48.5	36.3	42.1	—20.0	—62.6	—45.2
1934-35	100.5	92.9	96.0	88.6	99.2	93.5	—11.9	+ 6.3	— 2.5
1935-36	127.4	122.4	124.4	118.1	151.4	133.2	— 9.3	+ 29.0	+ 8.8
1936-37	128.7	92.6	106.8	177.2	88.8	137.9	+48.5	— 3.8	+31.1
1937-38	93.3	118.8	110.2	84.4	125.3	102.8	— 8.9	+ 6.5	— 7.4
1938-39	109.4	150.3	134.6	97.0	214.1	149.5	—12.4	+63.8	+14.9
1939-40	159.3	139.7	147.5	208.9	177.5	196.3	+ 49.6	+ 37.8	+ 48.8
1940-41	123.1	109.0	116.5	181.4	148.8	168.2	+58.3	+ 39.8	+ 51.7
1941-42	110.9	95.7	101.5	130.8	78.3	107.5	+19.9	—17.4	+ 6.0
1942-43	70.9	89.6	82.2	67.5	54.8	60.7	— 3.4	—34.8	—21.5
1943-44	78.1Inc	80.4	79.6	73.8	78.3	74.8	— 4.3	— 2.1	— 4.8
1944-45	138.7Inc	123.2Inc	129.7Inc	196.2	86.2	147.2	+ 57.5	—37.1	+ 17.5
1945-46	101.0Inc	101.0Inc	101.0Inc	120.3	107.0	114.5	+19.3	+ 6.0	+13.5
1946-47	76.4Inc	65.2Inc	69.3Inc	71.7	41 8	58.4	— 4.7	—23.4	—10.9
1947-48	147.9	98.1	117.7	130.8	65.3	102.8	—17.1	—32.8	—14.9
1948-49	63.2	71.6	68.3	38.9	62.7	49.1	—25.2	— 8.9	—19.2
1949-50	86.4	112.3	2.0	38.0	52.2	44.4	—48.4	—60.6	—57.6
			Harmony in percentage of times within 15 per cent:				40.0	37.0	47.0

TABLE 6

Further Variation Between Precipitation and Runoff in the Neckar Basin

Year	(Percentage of Normal) (a) In Upper Neckar Basin at Rottweil			(b) (Average of 3 precipitation stations) Between Winter Precipitation (Oct.-Mar.) and Summer Runoff (Apr.-Sept.)
	Oct.-Mar.	Apr.-Sept.	Ann.	
1920-21	+ 9.0	—40.7	—19.1	—18.0
1921-22	—38.6	+21.9	—17.2	+68.7
1922-23	+66.7	— 0.7	+43.7	—50.0
1923-24	—33.8	+59.5	+ 8.3	+67.5
1924-25	+15.3	—13.4	— 4.6	+20.7
1925-26	— 2.2	+14.1	+ 6.2	+18.3
1926-27	—30.2	+ 8.5	—15.0	+32.5
1927-28	+20.7	—17.4	+ 7.2	—40.7
1928-29	—39.8	—28.7	—34.9	—38.0
1929-30	—49.8	—16.0	—37.4	+ 2.3
1930-31	+ 1.0	+50.8	+25.1	+33.6
1931-32	+18.9	—37.0	+10.9	+34.4
1932-33	—25.5	—12.0	—21.7	+ 6.6
1933-34	—27.8	—74.6	—55.0	—32.2
1934-35	—14.3	+11.8	— 0.1	— 1.3
1935-36	—17.8	+23.0	+ 1.8	+24.0
1936-37	+42.3	— 9.7	+24.8	—39.9
1937-38	— 5.3	— 8.9	—13.6	+32.0
1938-39	+ 5.0	+52.1	+15.5	+104.7
1939-40	+57.6	+42.7	+54.9	+18.2
1940-41	+55.1	+48.4	+57.4	+25.7
1941-42	+35.9	—14.9	+13.7	—32.6
1942-43	— 9.1	—31.1	—21.5	—16.1
1943-44	— 6.6	+ 6.9	— 0.2	+ 0.2
1944-45	—	—	—	—52.5
1945-46	—	—	—	+ 6.0
1946-47	—	—	—	—34.6
1947-48	+18.3	—23.4	+ 4.6	—82.6
1948-49	—14.6	—10.6	—15.9	— 0.5
1949-50	—29.0	—53.8	—46.0	

(Percent of Normal)

Year	Oct.-Mar. Neckar	Oct.-Mar. Main	Apr.-Sept. Neckar	Apr.-Sept. Main	Annual Neckar	Annual Main	Harmony Oct.-Mar.	Harmony Apr.-Sept.	Harmony Annual
1920-21	46.4		28.7		37.4				
1921-22	63.3		161.9		107.5				
1922-23	181.4		67.9		130.8				
1923-24	99.2		182.8		135.5				
1924-25	75.9	97.8	86.2	90.7	79.4	95.4	21.9	4.5	16.0
1925-26	109.7	137.8	120.1	146.5	114.5	141.8	28.1	26.4	27,3
1926-27	75.9	93.3	134.1	165.1	102.8	121.8	17.4	31.0	19.0
1927-28	97.0	127.4	44.4	82.6	74.8	110.9	30.4	38.2	36.1
1928-29	50.6	63.7	54.8	62.8	51.4	63.6	13.1	8.0	12.2
1929-30	42.2	48.9	94.0	93.0	65.4	66.4	6.7	1.0	1.0
1930-31	145.6	155.6	177.5	118.6	158.9	141.8	10.0	58.9	17.1
1931-32	86.5	81.5	101.8	94.2	93.5	87.3	5.0	7.6	6.2
1932-33	52.7	78.5	80.9	83.7	65.4	80.9	25.8	2.8	15.5
1933-34	48.5	44.4	36.3	36.0	42.1	41.8	4.1	0.3	0.3
1934-35	88.6	80.0	99.2	103.5	93.5	89.1	8.6	4.3	4.4
1935-36	118.1	88.9	151.4	123.2	133.2	102.7	29.2	28.2	30.5
1936-37	177.2	157.0	88.8	109.3	137.9	139.1	20.2	20.5	1.2
1937-38	84.4	70.4	125.3	57.0	102.8	65.5	14.4	68.3	37.3
1938-39	97.0	87.4	214.1	74.4	149.5	82.7	9.6	139.7	66.8
1939-40	208.9	145.2	177.5	148.8	196.3	147.3	63.7	28.7	49.0
1940-41	181.4	143.0	148.8	180.2	168.2	158.2	38.4	31.4	10.0
1941-42	130.8	137.0	78.3	95.3	107.5	121.8	6.2	17.0	14.3
1942-43	67.5	63.0	54.8	60.5	60.7	61.8	4.5	5.7	1.1
1943-44	73.8	68.9	78.3	112.8	74.8	86.4	4.9	34.5	11.6
1944-45	196.2	130.4	86.2	109.3	147.2	122.7	65.8	23.1	24.5
1945-46	120.3	100.0	107.0	74.4	114.5	90.9	20.3	32.6	23.6
1946-47	71.7	78.5	41.8	60.5	58.4	71.8	6.8	18.7	13.4
1947-48	130.8	—	65.3	—	102.8		—	—	—
1948-49	38.0		62.7		49.1		Harmony within 15 percent — (Percent of total years)		
1949-50	38.0		52.2		44.4				
							52.0	35.0	48.0

Variation between Precipitation and Runoff

Harmony of precipitation does not necessarily imply similar harmony of precipitation and runoff. Thus the percentage of harmony within 15 percent between (1) winter precipitation and winter runoff, (2) winter precipitation and summer runoff, and (3) summer precipitation and summer runoff presents considerable variation:

1. Winter precipitation and winter runoff

Swiss Rhine	Neckar	Main	Moselle-Meurthe
—	40	40	42

2. Winter precipitation and summer runoff

Swiss Rhine	Neckar	Main	Moselle-Meurthe
54*	20	40	17

3. Summer precipitation and summer runoff

Swiss Rhine	Neckar	Main	Moselle-Meurthe
90	37	45	25

The detailed variations between precipitation and runoff by season and year in the Neckar Basin alone are given in Table 5. On the basis of seasons, the harmony in winter is only 40 percent and summer 37 percent. The entire year is 47 percent or less than half the years within the variation of 15 percent.

The harmony between runoff and precipitation at Rottweil in the upper Neckar Basin where the measured runoff occurs is still no better, being 33, 40 and 44 per cent for winter, summer, and annual; and the harmony between summer runoff and the previous winter precipitation for the entire Neckar Basin is only 20 percent or one year in five. Details are given in Table 6.

Indeed Professor Maurice Pardé of the Institut de Géographie Alpine, Grenoble, France writes that «Below Basel the Rhine Basin furnishes no more than a small quantity of melted snow. .. But nearly all this melt-water runs off in times of rising temperature and warm rains, in the course of the winter, or later at the beginning of April. Practically never is this yield appreciable in May. In the total from April to July, it is so small as to be negligible.» (Translation by Dr. S. B. Doten). Therefore where the efficiency of the stream is low, diversity between precipitation and runoff. is to be expected.

Harmony in Runoff between the Neckar and Main

In only about one-half of the 25 years of record are the Neckar and the Main within 15 percent of each other in either seasonal or annual flow. Indeed, in summer the proportion of years is only one third. See Table 7.

Although the Neckar has a stream efficiency of 46.2 percent and the Main only 29.7 percent, the harmony between precipitation and runoff in the Main seems to be steadier. For example,

(*) However, in the case of the accumulated snow cover as measured by snow survey, the harmony between snow and summer runoff is 75 per cent.

	(Percentage of times within 15 percent)		
	Winter Precip. to Winter Runoff	Summer Precip. to Summer Runoff	Winter Precip. to Summer Runoff
Neckar	33	40	20
Main	40	45	40

The presence of a snow cover, which is associated with efficiency of runoff, should be further studied on both streams.

The close harmony earlier noted between the Swiss Rhine and the Rhone has also been noticed by Ing. Aimé Coutagne, who suggests that a stream-forecast system would have even larger practical value on the Po of Italy.

Nevada (U. S. A.) Agricultural Experiment Station, Reno.

COMME ON MESURE LA NEIGE EN ITALIE

Dott. Ing. MASSIMO LEVI
Directeur du Bureau Hydrographique du Magistrato alle Acque — Venise

Prémisses

Dans le siècle passé les mesurages de la quantité de pluie tombée, avec d'autres observations à caractère purement météorologique, étaient effectués par les observatoires astronomiques (1), tandis qu'on ne mesurait pas la quantité de neige tombée.

La nécessité et l'importance d'étudier les précipitations de neige sont devenues évidentes lorsqu'on commença les grands travaux hydrauliques.

En effet l'étude d'un bassin hydrologique ne peut pas être complète si, outre les données de pluie d'une période assez longue, on ne connait aussi la quantité de neige qui peut s'accumuler sur le bassin même, pendant l'hiver.

La determination de la consistence de la neige precipitée a une importance spéciale en Italie, dans les bassins où la surface des zones à cote élevée peut être très considérable par rapport à la surface totale; comme, par exemple, dans le bassin du fleuve Adige, qui, après le Pô, est le plus important de l'Italie septentrionale et qui a une surface de 11.954 Km^2. dont 4.617 Km^2 (c'est à dire le 39%) à cote superieure à 1.800 m. avec 212 Km^2 de glaciers.

Il est évident que pour l'étude des neiges d'un bassin hydrologique particulier on doit connaître très bien les caractéristiques morphologiques du bassin même, c'est à dire l'orientement, l'altitude moyenne et la végétation (en presence de bois la neige se conserve plus longuement).

En Italie le recueil et l'élaboration des données des précipitations neigeuses ont été commencés à peu près en 1907, dans le bassin du Pô et dans la Vénétie, tandis que pour le restant de l'Italie, ils ont été commencés en 1917 avec l'institution du « Service Hydrographique ».

Observations et Mesurage des Précipitations Neigeuses

Les observation sur la quantité de neige tombée sont effectuées dans les stations mêmes du Service Hydrographique où les pluies sont relevées.

Les observateurs enregistrent chaque chute de neige et déterminent, avec la fonte, l'équivalent en eau de la neige reçue par le pluviomètre et mesurent tous les jours l'épaisseur de la neige sur le terrain.

Dans les stations d'observation pourvues d'instrument enregistreur (pluviographe) on relève tout de suite l'équivalent en eau de la neige tombée, lorsque il a été possible de pourvoir la petite cabane, où l'instrument est installé, d'une résistance électrique (dont la chaleur provoque la fonte de la neige).

Pour déterminer l'épaisseur de la neige les stations sont pourvues d'une perche nivômétrique de hauteur convenable, avec des graduations de 5 en 5 centimètres, plantée en un endroit convenable, qui ne soit pas exposé aux vents.

La consistance numérique du réseau pluviométrique italien est indiquée dans le tableau suivant, où l'on peut voir aussi le nombre des stations pourvues de ther-

(1) L'observatoire astronomique de Padoue, le plus ancien d'Italie, commença ses observations dès 1713.

momètres à maxima et minima : les données thermométriques sont très utiles pour l'étude des conditions de la couche de neige.

Type du réseau d'observations	Distribution des stations en cote de m.								Total
	0.250	250 500	500 750	750 1000	1000 1500	1500 2000	2000 2500	outre 2500	
Pluviométrique	1217	890	639	375	308	80	29	12	3550
Thermométrique	548	159	61	51	47	5	—	—	871

Fig. 1 — Station pluviométrique de Vernago (Adige) à la cote 1700, comprenant un pluviographe, un thermographe et une perche pour mesurer la hauteur de la neige.

Dans le même tableau on relève l'existence de stations d'observation à des cotes superieures à 2500 m.

Dans ces stations sont installés des instruments capables de recevoir toutes les précipitations (pluvionivomètres totalisateurs) : un mélange spécial qui empêche la congélation provoque la fonte de la neige reçue et une couche d'huile de vaseline empêche l'évaporation. Ces instruments sont installés près des refuges alpins, et avec eux, de temps à autre, même pendant l'hiver on effectue des observations.

Mesurage de la couche Neigeuse et son équivalent en Eau

Le mesurage de l'épaisseur de la neige sur le terrain (et de son équivalent en eau) n'est pas particulièrement difficile dans les stations d'observation stables, parce

Fig. 2 — Station pluvio-nivométrique totalisatrice à la cote 2950 Refuge Bellavista (Val de Senales — Adige).

qu'il suffit que près de chaque station soit installé, dans un endroit convenable, la perche métrique dont nous avons déjà parlé.

La détermination de l'équivalent en eau de la couche neiguese, au contraire, n'est pas aussi simple dans les stations provisoires opportunément choisies sur les points les plus caractéristiques d'un bassin hydrologique.

Cette determination est nécessaire pour réunir les élements suffisants pour faire, dans un moment donné, le calcul, le plus approximatif possible, de la quantité d'eau dont on peut disposer dans le bassin d'après la présence de la couche neigeuse.

Parmi les méthodes suggérées par plusieurs experts la plus précise et la plus pratique a été retenue, celle qui emploie le « Pèse-neige ».

Naturellement, pour obtenir des résultats avec une bonne approximation et convenant aux fins pratiques de cette recherche, le pèse-neige doit garantir que, dans son emploi, des eventuelles erreurs de construction, ou autres, n'infirment pas les résultats des observations.

A ce propos, il faut rappeler les approximations auxquelles on arrive en météorologie avec l'emploi des instruments suivants dont on fait usage pour le mesurage des précipitations :

— le pluviomètre, avec une embouchure de 0.10 m² généralement employé en Italie, qui garantit le mesurage de la pluie, jusq'au dixième de millimètre;
— le pluviographe, qui donne une approximation de 0.2 mm.;
— le pluvio-nivomètre totalisateur, qui donne une approximation de 2 mm.

Le pèse-neige employé par le « Service Hydrographique Italien » est basé sur les études, qui seront ici resumées, du Professeur Giuseppe Crestani (1) faites à l'aide du pèse-neige type « T.70 » fabriqué depuis 1921 par le « Bureau Hydrographique du Magistrato alle Acque » de Venise, dont on omit la description particularisée parce qu'il est analogue au type « C.N.1 » qu'on décrira par la suite.

(2) G. Crestani — « Sur la détermination de l'équivalent en eau et de la densité de la neige » — Étude théorique et expérimentale — Extrait des Annales des Travaux Publiques (ancien Journal du Génie Civil) année 1937 — fascicule N° 6.

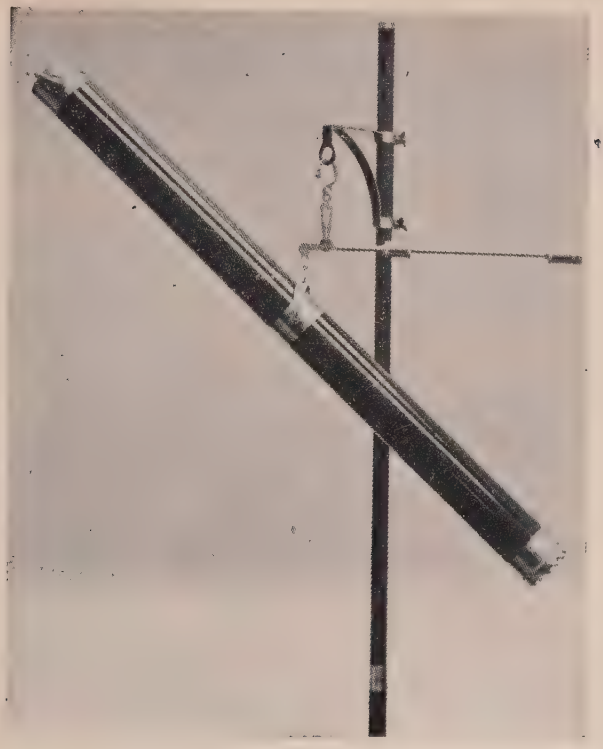


Fig. 3 — Pèse-neige « T. 70 ».

Il est constitué essentiellement par un tube cylindrique muni à l'extrémité inférieure, d'une couronne à couteau coupe-neige et d'une balance romaine (plus avantageuse, quoique plus lourde, qu'une balance à ressort).

Il faut dire d'abord que, pour avoir l'équivalent en eau à l'aide du pèse-neige susdit, on prélève et on pèse un spécimen de neige en guise de cylindre, ayant une section de surface « S» .

Le poids « P », exprimé en grammes, divisé par la surface « S » exprimée en cm^2, donne l'équivalent en eau, c'est à dire l'hauteur en centimètres de l'eau qui résulterait de la fonte d'une quantité de neige égale au prélèvement contenu dans un récipient cylindrique de section « S ».

Le rapport entre cet équivalent (u) et l'épaisseur (v) donne la densité; le contraire est appelé « hauteur spécifique ».

En effet les résultats óbtenus à l'aide du pèse neige sont rendus moins précis, outre que par l'instrument, aussi par celles relatives aux possibles fausses manœuvres, déterminées par la diversité de l'épaisseur. de la densité et de la constitution de la neige.

Il est donc évident que les opérations de prélèvement doivent être faites d'une façon systématique et rationnelle.

L'approximation de 0.5 cm. dans le mesurage de l'épaissseur de la neige, le diamètre du couteau coupe-neige (70 mm) avec l'approximation de $\pm$ 0.1 cm., et la sensibilité de 4 grammes de la balance sont des conditions qu'on obtient facilement. Elles sont suffisantes pour donner des résultats assez exacts, aux buts de la recherche, dans les calculs de l'équivalent en eau et de la densité de la neige.

Quant à l'équivalent en eau (u) de la neige, étant $u = \frac{P}{S}$, l'erreur absolue δu, qui se réflechit sur « u » en conséquence des erreurs δP, δS dans le mesurage de P et de S, est:

$$1)\ \delta u = \frac{\delta P}{S} - \frac{P}{S} \cdot \frac{\delta S}{S}\ ;$$

et l'erreur relative est :

$$2)\ \frac{\delta u}{u} = \frac{\delta P}{P} - \frac{\delta S}{S}\ ;$$

indiquant avec « d » le diamètre et avec δd son erreur, on a :

$$\frac{\delta S}{S} = \frac{2\,\delta d}{d} = \frac{0.02}{7} = 0.003,$$

où, comme erreur possible du diamètre, on a pris un dixième de millimètre.

Pourtant, lorsque l'erreur du diamètre est égale ou inférieure a 1/10 de mm., l'erreur pour cent qui à cause de cela se réfléchit dans le calcul de l'équivalent est égale ou inférieure a 3/1000.

Telle erreur est constante pour un instrument donné; il s'en suit que par le vérification expérimentale on a eu la certitude que δu est inférieur à 0.1 mm. (comme il se vérifie en réalité); par conséquent il ne vaut pas la peine de calculer δS, qu'on peut, étant donné sa petite importance, négliger.

Pour déterminer l'erreur de la pesée on emploie aussi l'équation (1) qui donne l'erreur absolue.

Puisque le deuxième terme est le produit $u\delta S$, où, étant donné qu'il s'agit de neige, « u » est toujours moindre que 0.5 et pourtant négligeable, elle devient :

$$3)\ \delta u = \frac{\delta P}{S}$$

La balance romaine, fabriquée exprès, a une sensibilité bien plus grande que 4 grammes, mais, en admettant aussi l'erreur la plus grande de 4 grammes, de la 3) on obtient que l'erreur absolue qui s'en suit dans l'équivalent, quand la section « S » du tube est à peu près de 40 cm^2., est de 1 mm.

Cette approximation de 1 mm. est plus que suffisante, car il s'agit de mesurer des précipitations, où il y a souvent des erreurs et des incertitudes bien plus grandes.

Puisqu'on a démontré que l'erreur de l'équivalent en eau est due à la balance, il n'y a pas de raison d'employer des tubes de diamètre supérieur a 70 mm.

En augmentant le diamètre susdit on aurait plus de difficultés de pénétrer dans la neige dure et on rendrait l'instrument plus encombrant; au contraire une sensible diminution du diamètre au dessus des 70 mm. tandis qu'elle faciliterait la pénétration dans la couche de neige, ferait trop sentir la présence dans la neige d'éventuels éléments étrangers et exigerait aussi une plus grande sensibilité de la balance.

On peut donc retenir les erreurs maxima de « u » et de « v » :

$$\delta u = \pm 0.1 \text{ cm.}; \delta v = \pm 0.5$$

Puisqu'on a : $\delta \left(\frac{u}{v}\right) = \frac{\delta u}{v} - \frac{u}{v}\frac{\delta v}{v}$.

l'erreur maximum ($y = \frac{u}{v}$) dans la densité ($x = \frac{u}{v}$), en prenant les valeurs absolues des erreurs et leur somme, devient :

$$3') \; y = \frac{u}{v}(0.1 + 0.5\, x).$$

Si x est constante, l'erreur absolue y dans le calcul de la densité est plus petite autant que plus grand c'est l'épaisseur v, et précisement la relation entre l'erreur y et l'épaisseur v, aux diverses densités x, est donnée par les hyperboles équilatérales (fig. 4) définies par l'équation :

$$4) \; yv = 0.1 + 0.5\, x$$

En indiquant avec $z = \frac{y}{x}$ l'erreur pour cent de la densité de la précédente on a :

$$5) \; zv = \frac{0.1}{x} + 0.5$$

et celle-ci nous donne la relation entre l'erreur pour cent de la densité et l'épaisseur v du spécimen prélevé; relation qui exprime la constance du produit des deux grandeurs pour chaque valeur de x (fig. 5).

Tandis que le produit de l'épaisseur pour l'erreur absolue augmente avec l'augmentation de la densité, le produit de l'épaisseur pour l'erreur pour cent augmente avec la diminution de la densité.

Les considérations précédemment faites et de l'examen des graphiques (fig. 4-5) qui représentent les équations 4) et 5), on peut ainsi conclure :

1) l'erreur qu'on fait dans la détermination de l'équivalent en eau de la neige est essentiellement due à la pesée;
2) Le diamètre du tube qui le mieux répond aux exigences pratiques et théoriques est à peu près de 70 millimètres;
3) L'erreur absolue, qui, en conséquence des erreurs dans la détermination de l'épaisseur de la neige et de l'équivalent en eau, s'étend au calcul de la densité, à égalité de densité, diminue avec l'accroissement de l'épaisseur et, à égalité d'épaisseur, (augmente avec l'accroissement de la densité; en outre elle est inférieure à 0.01 c'est-à-dire qu'elle est inférieure à une unité du deuxième chiffre décimal) :

avec densité 0.1 pour épaisseurs $>$ 15 cm.
» » 0.2 » » $>$ 20 »
» » 0.5 » » $>$ 35 »

L'erreur pour cent diminue aussi avec l'accroissement de l'épaisseur trouvé, mais, à différence de l'erreur absolue, augmente avec la diminution de la densité.

Pour des épaisseurs supérieures à 30 cm., elle reste inférieure au 5 % avec densité égale à 0.1; avec des densités plus importantes elle est inférieure à 3.3 %.

Pourtant, s'il s'agit d'épaisseurs supérieures à 30 cm., ce qui arrrive d'habitude, l'erreur est inférieure au 5 %, quelle que soit la densité.

Dans ces dernières années, les études relatives à la connaissance des précipitations neigeuses ont été conduites avec beaucoup de promtitude par la Commission des Neiges du Comité Italien de Glaciologie, laquelle, en profitant des résultats obtenus à l'aide du pèse-neige « T.70 », a étudié un nouveau type de cet instrument, moins lourd et plus maniable.

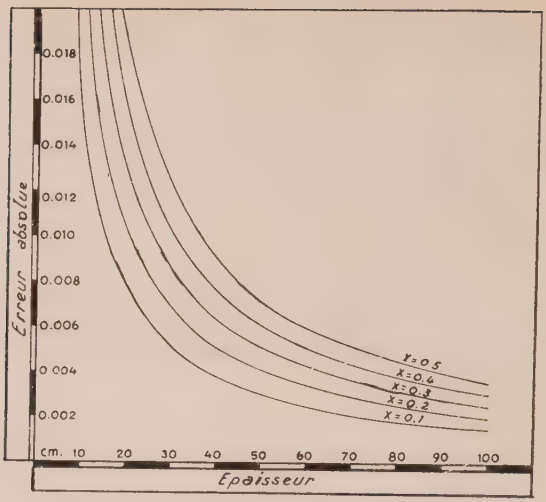


Fig. 4

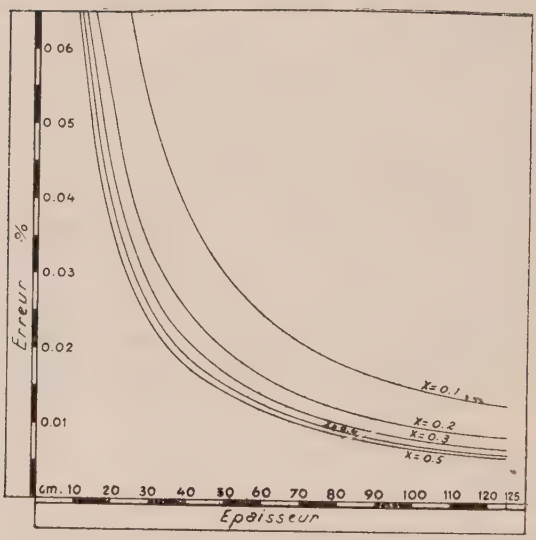


Fig. 5

Il est employé dans des stations d'observation et d'étude nommées « Stations pilotes » situées en zone et bassins caractéristiques de haute montagne, opportunément choisies des Alpes et des Apennins.

Le nouveau type de pèse-neige, nommé « C.N. 1 » a été adopté aussi par le « Service Hydrographique Italien ».

Dans chaque « Station — pilote » pourvue d'instruments pour les observations météorologiques, les mesurages des neiges à l'aide de l'instrument susdit, sont faites

systématiquement sur le même point-base; de cette façon on peut étudier les mêmes mesurages en rapport avec les facteurs météorologiques du lieu.

En outre, pour chaque « Station-pilote», on choisit préventivement des stations auxiliaires.

Ces stations auxiliaires sont placées dans des localités caractéristiques du territoire de compétence, avec expositions diverses, où l'on peut faire des relèvements pour connaître avec suffisante approximation l'épaisseur et l'équivalent en eau de la couche de neige de cette zone.

Ces localités doivent être facilement reconnaissables, par des releveurs-skieurs, par les caractéristiques du terrain et par des signaux opportunément prédisposés.

En outre elles doivent être situées à des cotes diverses, le long d'un où de plusieurs itinéraires à parcourir à l'aide des skis dans une période de temps relativement courte afin d'éviter dans cette même période l'influence dûe à des variations météorologiques sensibles. Le long de ces itinéraires ou peut aussi établir, toutes les fois, des autres stations provisoires.

Avec toutes les mesures prises, et surtout avec celles effectuées après l'hiver et avant le dégel de printemps, ou peut obtenir des données assez exactes sur la quantité de neige accumulée dans le bassin de compétence de la « Station-Pilote ».

De cette façon on détermine le volume d'eau correspondant, dont ou peut disposer, outre à ceux dérivants des pluies et des éventuels apports de glaciers, pour l'utilisation.

Description du Pése-Neige Type « C.N.1 »

Dans le nouveau type de pèse-neige, avec balance romaine, usé pour l'étude de la couche de neige dans les « Stations-Pilote » à la place des plusieurs types employés jusqu'à maintenant, le diamètre du couteau coupe-neige est encore de 70 mm.

Pour rendre plus facile son transport en haute montagne par les operateurs-skieurs, il est constitué, au lieu d'un seul tube en fer de la longueur de 1 m. où 2 m., par 4 élements tubulaires en duraluminium (fig. 6 *b*-C) chacun de la longueur de 50 cm., qu'on met dans un sac fait exprès pour cela avec la balance et les autres accessoires.

Des 4 tubes, numérotés de 1 à 4, qui constituent le pèse-neige, on emploie un ou plusieurs éléments, par rapport à l'épaisseur de la couche de neige.

Chacun des 4 susdits éléments du tube a extérieurement une graduation de 0 à 50 cm., ce qui permet de mesurer exactement la hauteur de la neige dans le point où l'on fait le prélèvement.

A l'extrémité supérieure du premier élément est vissée transversalement une poignée, qu'on use pour la manœuvre rotatoire d'enfoncement du tube, pendant le prélèvement, tandis que dans la partie inférieure on peut visser selon les cas un ou plusieurs éléments du tube, ou bien directement le couteau coupe-neige.

Le pèse-neige « C.N.1 » est pourvu de trois divers types de couteau coupe-neige, avec couronne dentée, qu'on visse, à l'aide d'une clef faite exprès (fig. 6-*g*), à l'extrétrémité du tube, selon la qualité de la neige à prélever, et précisément :

— *Couteau n.* 1 (fig. 6 *d*), consitué par un simple couteau avec une couronne dentée, pour l'emploi en présence de *neige compacte* [3] et particulièrement convenable pour les prélèvements horizontaux en tranchée;

— *Couteau n.* 2 (fig. 6 *e*), constitué par un couteau avec couronne dentée à l'inté-

[3] Symboles *d*) et *e*) de la classification internationale de la neige;

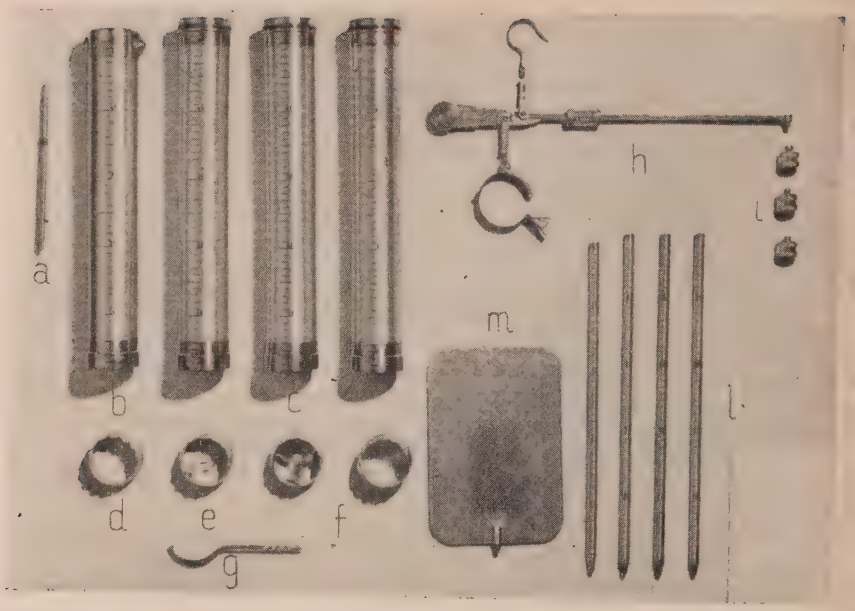


Fig. 6

rieur duquel est fixée une lame hélicoïdale disposée diamétralement, pour faciliter l'entrée de la neige dans le tube, qu'on emploie en présence de neige de moyenne compacité (4)

— *Couteau n.* 3 (fig. 6 *f*), consittué par un couteau pourvu à l'intérieur, à 2 cm. du bord, de trois ailettes à charnière, avec la pointe légèrement pliée vers le centre du tube, lesquelles se soulèvent, en s'appuyant à la parois du couteau, pour la pression de la neige pendant l'enfoncement, en se plaçant au contraire horizontalement pendant l'extraction du tube même, toutes les fois que la neige tend à sortir. Le type de couteau est particulièrement utile pour prélever la *neige farineuse et sèche* (5).

Pour faciliter le remplissage du tube avec la neige pendant l'enfoncement, la section du tube même est légèrement plus petite que celle du couteau; pour vider l'instrument après l'emploi, on le renverse après en avoir enlevé la poignée fixée dans la partie supérieure.

Pour la mesure préliminaire de l'épaisseur de la couche de neige, le pèse-neige est pourvu d'un double mètre centimètré (fig. 6 *l*), constitué par 4 éléments égaux, vissables, en tube d'aluminium.

Ces éléments, unis où séparés, peuvent servir de manche pour une palette (fig. 6 *m*) elle aussi en aluminium, de la dimension 20 × 30 centimètres en dotation du pèse-neige, dont on se sert pour prélever les spécimens horizontaux et pour les

(4) Symbole *d*) de la classification internationale de la neige.
(5) Symboles *a*) et *b*) de la classification internationale de la neige.

Fig. 6 (*bis*)

spécimens verticaux supérieurs à deux mètres. La palette limite exactement la hauteur du cylindre de neige qu'on prélève, en evitant des éventuels déplacements de la masse de neige dûs à la compression vers l'intérieur.

Dans la balance romaine (fig. 6 *h*) la graduation de la perche va de 0 à 2 Kg.; elle donne le poids total effectif de l'instrument et de la neige prélevée.

La graduation permet la lecture immédiate de 10 gr. en 10 gr., toutefois le peson glissant porte un nonius, à l'aide duquel on peut obtenir l'approximation d'un gramme; la balance est pourvue aussi d'un collier dénoué, qui permet de suspendre opportunément le pèse-neige.

Pour conserver à la balance l'approximation susdite, on a bâti trois poids-lest égaux (fig. 6 *i*) à chacun desquels correspond un équivalent de 2 Kg. Ces poids, sont, selon les nécessités, vissés à l'extrémité de la perche graduée de la balance, en permettant une portée utile maxima de 8 Kg., suffisante pour le fonctionnement du pèse-neige complet de tous les quatre tubes.

Modalités pour l'emploie du Pèse-Neige

Pour effectuer le mesurage de la neige on enfonce le tube de prise dans la couche de neige, et on lui donne un léger mouvement rotatoire, jusqu'à quand on arrive au terrain, en cas de prélèvement vertical, ou à la palette,en cas de prélèvement horizontal. De cette façon un cylindre de neige de diamètre égal à celui du couteau entre et reste dans le tube.

Après avoir effectué cette opération on attache le tube, à l'aide du collier denodé fait exprès, à la balance et on fait la pesée (fig. 6 bis).

On obtient le poids net de la neige contenue dans le tube en faisant la soustraction de la tare du poids brut; la tare est préventivement relevée avec la même balance ou bien lue sur le tableau préparé exprès et dont tous les instruments sont pourvus.

Pour être sûr de l'exactitude du prélèvement exécuté, on vérifie, avec le double mètre en dotation, la profondeur atteinte par le spécimen après l'extraction du pèse-neige, afin de s'assurer qu'une partie du cylindre de neige ne soit pas restée sur le terrain. En ce cas on doit refaire le prélèvement en localité proche et intacte de la couche de neige.

Lorsqu'on est en présence de neige légère et farineuse on emploie le couteau n. 3, avec les ailettes à charnière, et on doit se rappeler que, quand les ailettes sont baissées, une couche de neige d'environ 5 cm. sort du tube et pour cela n'est pas pesée, tandis que la même couche de 5 cm. est comprise dans l'épaisseur de la couche de neige mesurée précédemment.

Opérations de Prélèvement

Pour obtenir des résultats vraiment sûrs il faut :

a) vérifier la balance du pèse-neige, et particulièrement sa senbilbilité;

b) choisir le couteau coupe-neige par rapport au type de neige :

— pour neige compacte couteau n. 1 ([6])
— pour neige de moyenne consistence » »n. 2 ([7])
— pour neige farineuse » n. 3 ([8])

En outre il est nécessaire que la zone du prélèvement, normalement connue à terrain découvert, ne soit pas en proximité de sentiers, pour éviter le piétinement de la neige; on doit éviter les zones battues par les forts vents dont l'action modifie la distribution de la couche de neige de façon à la rendre très irrégulière.

Afin de donner aux mesures le juste poids, pour chaque prélèvement, ou groupe de prélèvements dans une zone déterminée, on doit indiquer :

1) L'épaisseur de la couche de neige;
2) La température de l'air du lieu mesurée à peu près à 1.50 m. au dessus du niveau de la neige, et de la neige à sa surface;
3) L'exposition de la zone (au soleil ou à l'ombre, au nord ou au sud, etc.);
4) La situation du terrain (plan, en légère pente, à forte pente);
5) La direction et le type du vent (faible, modéré, fort ou à rafales);
6) L'état du ciel (serein, couvert, à moitié couvert).

La température de l'air doit être prise à l'ombre, eventuellement aussi à l'ombre de l'observateur même.

On relève la température de la neige à la surface en plongeant le bulbe du thermomètre dans la neige jusqu'à la couverture totale et en faisant la lecture après 15'.

([6]) Symboles *d*) et *e*) de la classification internationale de la neige
([7]) Symbole *d*) de la classification internationale de la neige.
([8]) Symboles *a*) et *b*) de la classification internationale de la neige.

a) *Prélèvements verticaux* (fig. 7) : avant tout on s'assure de l'épaisseur de la couche de neige, à l'aide du double mètre en aluminium fait exprès, pour savoir de combien on est près du terrain et ón compose le tube avec le nombre d'éléments suffisant par rapport à l'épaisseur verifié.

Après on applique le couteau coupe-neige convenable au type de neige, on fait le prélèvement du spécimen en enfonçant le tube dans la neige et en donnant à la poignée une pression constante et un mouvement continuel de rotation jusqu'à rejoindre le terrain. Après cela, pour contrôler l'operation on fait sur la graduation extérieure du pèse-neige la lecture relative à l'épaisseur, on extrait le tube et on l'attache à la balance.

Avant de faire la pesée il faut avoir soin d'enlever du couteau les fargments de terrain qui se sont éventuellement attachés.

Par la suite il faut contrôler si le prélèvement est complet.

Fig. 7

Fig. 8

On fait la pesée, on vide le tube en le renversant, ayant précédemment enlevé la poignée.

Il faut enfin s'assurer que dans le spécimen prélevé ne se soient pas mêlées à la neige des matières étrangères (pierres, herbes, rameaux etc.); en tel cas il faut répéter le prélèvement.

b) *Prélèvement horizontaux* (fig.8) : pour effectuer les prélèvements horizontaux il est nécessaire de creuser dans la couche de neige une tranchée aux dimensions suffisantes pour pouvoir bien manœuvrer le tube n. 1, c'est à dire celui pourvu de poignée.

On doit creuser la tranchée avec la précaution de ne pas piétiner la couche de neige et d'arranger la neige de rebut latéralement à la parois où l'on fait le prélèvement.

Pour exécuter le prélèvement il faut employer la palette qu'on enfonce dans la couche de neige à une distance de 0.50 m. du bord de la parois de la tranchée.

On doit exécuter les prélèvement horizontaux dans chaque couche de neige qui par ses caractéristiques diffère de la couche successive, exception faite pour les couches d'épaisseur inférieure au diamètre du pèse-neige; dans ce cas il sera nécessaire de faire le prélèvement en cherchant d'obtenir un spécimen moyen.

ТЕТРАГОНАЛЬНЫЕ СИСТЕМЫ ЛЕДЯНЫХ ЖИЛ

(Теория образования, палеогеографическое и геологическое значение)

Б. Н. Достовалов.

§ 1. Введение

Полигональные системы ледяных жил широко распространены в суглинках и торфяниках Арктики и Субарктики. Они представляют подавляющую часть крупных масс подземных льдов, иногда составляют свыше 50 % объема верхнего тридцатиметрового слоя породы, часто придают специфический характер ландшафту и обладают сложными структурой и формой, отражающими фациальные условия и историю их развития (Рис. 1).

Ледяные жилы растут на основе повторных морозобойных растрескиваний мерзлых пород и цементаций трещин льдом, поэтому выяснение и учет закономерностей образования температурных трещин необходимы для понимания механизма их формирования.

В предлагаемой работе развитие жильных льдов рассматривается с точки зрения закономерностей образования температурных трещин в сочетании с теплофизическими условиями и процессом накопления осадков. Полученные закономерности объясняют наблюдаемую форму залегания жильных льдов и позволяют сделать некоторые палеогеографические и геологические выводы.

§ 2. Закономерности температурного растрескивания мерзлых пород и условия развития жильных льдов

В работе автора « О физических условиях образования морозобойных трещин и развития трещинных льдов в рыхлых мерзлых породах » (Б. Н. Достовалов, 1952) были определены температурные разрывающие напряжения в массиве, обладающем горизонтальной и вертикальной поверхностями и охлаждающемся сверху. Эти разрывающие напряжения (τ_b) равны

$$\tau_b = \frac{1}{2} G \alpha x \frac{\Delta t}{\Delta z} \qquad (1)$$

где G — модуль сдвига, α — коэффициент линейного расширения (сокращения), x — расстояние от свободной вертикальной поверхности, $\Delta t/\Delta z$ — градиент температуры по вертикали.

Из этой формулы вытекают следующие выводы :

1. Так как разрывающие напряжения пропорциональны расстояниям (x) от свободной вертикальной поверхности, в *однородных массивах должна образовываться система трещин, параллельных этой поверхности и друг другу*.

2. Так как вектор градиента температуры перпендикулярен изотермическим поверхностям, а последние параллельны свободным горизонтальным и вертикальным поверхностям массива, то система полос, выделяемая системой параллельных трещин, должна разбиваться поперечными трещинами на систему прямоугольных призм. *Углы при сопряжении продольных и поперечных трещин должны быть прямые.*

3. Разрывающие напряжения (τ_b) пропорциональны произведению линейного размера отдельности (x) на градиент температуры ($\Delta t/\Delta z$), поэтому *при малых градиентах однородный массив распадается на крупные прямоугольные отдельности, а затем при увеличении градиентов эти отдельности последовательно делятся пополам трещинами последовательных порядков генерации на все более мелкие отдельности* (рис. 2).

4. При сопряжении трещины в виде основания и перпендикуляра *основание образуется ранее трещиной более низкого порядка, а перпендикуляр — позже трещиной более высокого порядка.*

5. В неоднородных породах с переменными G и α трещины должны быть извилистыми и не вполне параллельными друг другу, но перпендикулярность в точке сопряжения остается в силе.

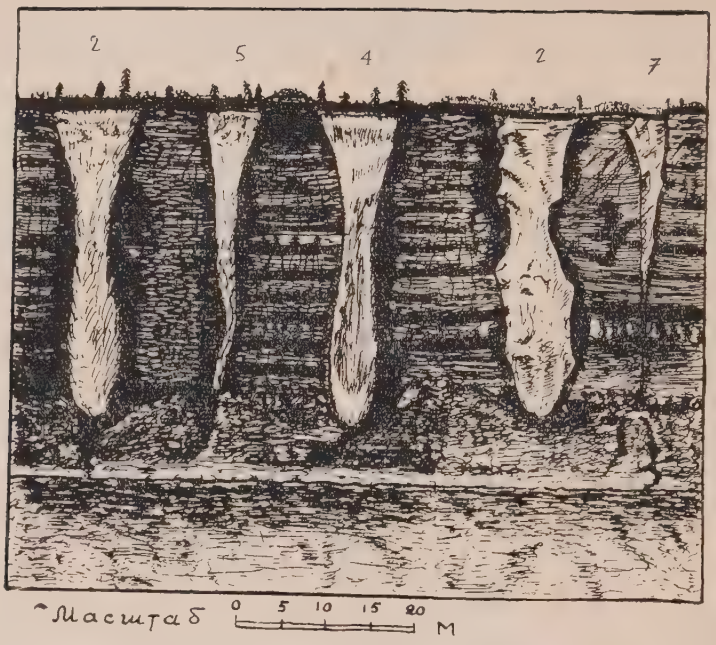


Рис. 1

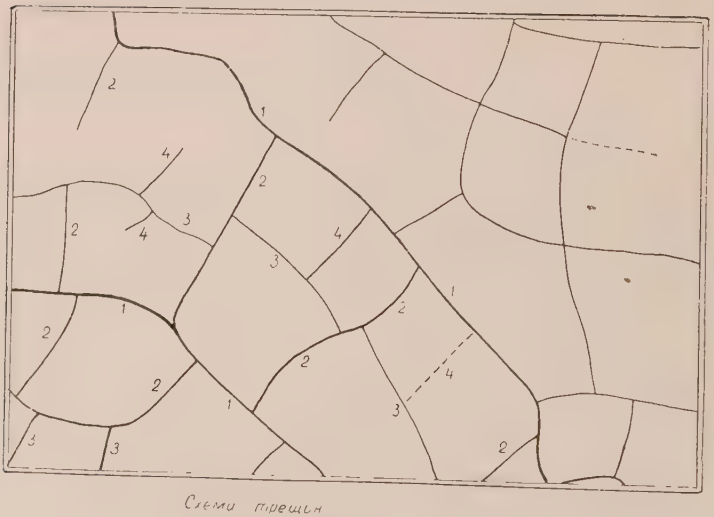


Рис. 2

6. В литологически различных породах (различные G и α) должны вообще развиваться системы отдельностей различной крупности.

7. Вследствие охлаждения свободных вертикальных поверхностей выделившихся отдельностей последние должны разбиваться горизонтальными трещинами на параллелепипедальные отдельности.

Так как тетрагональные жильные льды растут только в вечномерзлой толще при растрескивании и цементации последней, можно указать следующие условия их образования :

1. Температурные трещины должны проникать в вечную мерзлоту.
2. Должна обеспечиваться возможность цементации трещин льдом; если они становятся слишком узкими, рост прекращается.
3. Процесс растрескивания и цементации должен периодически повторяться в том же месте.
4. Порода должна представлять собой достаточно крупный массив.
5. Процессы протаивания не должны преобладать над процессами промерзания.

При несоблюдении хотя бы одного из этих условий тетрагональные системы ледяных жил не развиваются. Указанные условия их формирования приурочены к поймам рек, протекающих в породах с температурой ниже —3° С, возможны на побережьях арктических морей, а также повсюду, где накапливается элювий и имеется достаточное увлажнение и низкие температуры грунта.

§ 3. Элементарный цикл развития жильных льдов и вероятность повторного растрескивания

Морозобойное растрескивание впервые начинается зимой на участках пород, выходящих на поверхность вследствие отступания какого-либо водоема. Вода, попадающая в трещины, следующей весной во время половодья быстро замерзает и цементирует растрескавшуюся мерзлую толщу в сплошной массив. Кроме того, на площадях между трещинами в это время отлагается некоторый слой осадка.

Замерзая в трещинах, вода расширяется и давит на вмещающую породу, уменьшая в ней растягивающие усилия. Летом при повышении температуры верхнего слоя мерзлой толщи порода расширяется и растяжение сменяется сжатием. Последнее усиливается вследствие того, что трещина расклинена элементарной жилкой льда. Это расклинивание создаст новые условия при смене летнего сжатия растяжением при понижении температуры следующей зимой, так как последнее должно сначала снять увеличенные напряжения сжатия и только после этого может привести к новому уменьшению объема, растяжению и растрескиванию мерзлой породы.

Сочетание единичного растрескивания и заполнения трещины льдом образует элементарный цикл развития ледяной жилы, представляющей собой сумму большого числа периодически образующихся элементарных жилок.

При этом важно иметь в виду, что *вероятность повторного образования трещин одного и того же порядка уменьшается вследствие эффекта расклинивания и кроме того вероятность повторного растрескивания тем меньше, чем выше порядок генерации трещин*. В относительно теплые зимы градиенты температур малы и образуются трещины невысоких порядков, в суровые зимы градиенты велики и растрескивание доходит до трещин высоких порядков (рис. 2). Таким образом, элементарные жилки, приуроченные к трещинам высоких порядков, образуются не каждый год и тем реже, чем выше их порядок.

Поэтому поперечная мощность жил, растущих одновременно, но формируемых трещинами различных порядков, должна быть тем меньше, чем выше порядок формирующих их трещин (рис. 3).

§ 4. Толщина жил в зависимости от глубины трещин, ширины трещин и скорости накопления осадков

После первого растрескивания породы и цемнтации трещин льдом дальнейший рост ледяной жилы и форма ее поперечного сечения зависит от соотно-

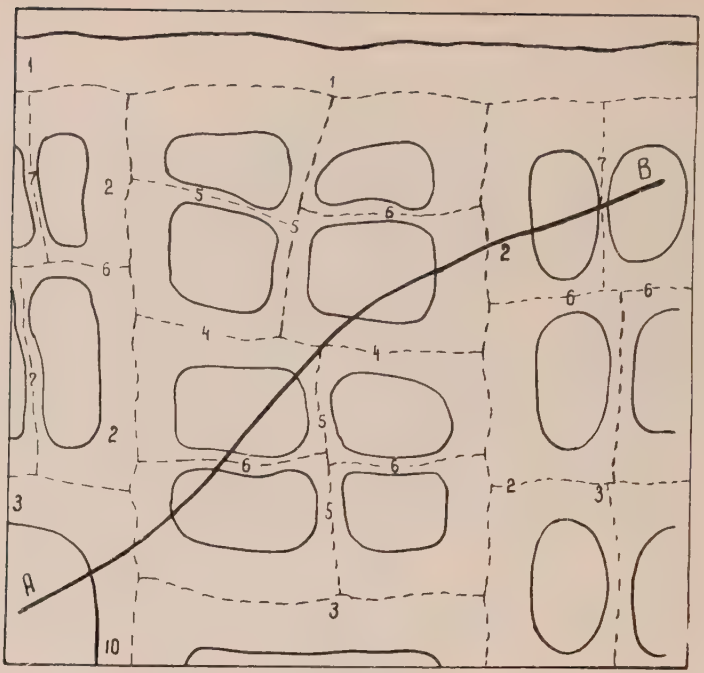


Рис. 3

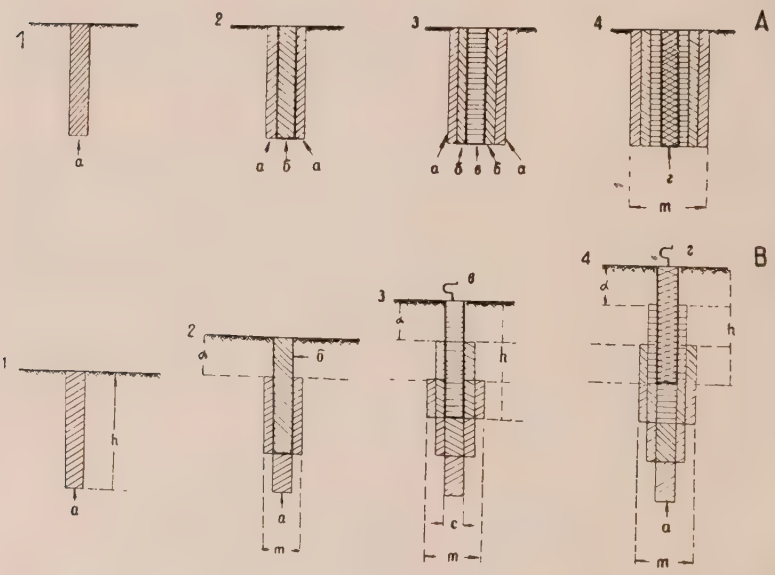


Рис. 4

шения между глубиной элементарной трещины (h), шириной повторных трещин (c) и скоростью накопления осадков (d).

Чтобы выяснить как рост жилы идет при различных скоростях накопления осадков, рассмотрим два крайних случая : (А) — осадки не накапливаются — эпигенетический рост и (В) — осадки быстро накапливаются — сингенетический рост (рис. 4).

В случае (А) элементарные жилки « a », « b », « c » и « d » последовательно вклиниваются друг в друга и рост жилы происходит только в толщину, которая равна сумме средних толщин элементарных жилок. В результате роста жила превращается в широкий клин, вертикальная мощность которого равна глубине растрескивания (h). Последовательные трещины проходят по льду, а не в породе, потому что прочность льда на разрыв меньше прочности мерзлой вмещающей породы. Вследствие накопления расклинивающего действия эпигенетический рост жил должен прекращаться.

При сингенетическом росте жил (случай В, рис. 4) и постоянной глубине растрескивания (h) каждая последующая элементарная жилка располагается выше предыдущей на величину d (скорость отложения осадков) и вновь образующиеся жилки совсем выходят за пределы горизонта 1 после h/d циклов. В этом случае рост жилы в пределах горизонта 1 прекращается, и она достигает предельной толщины m. Эта толщина очевидно равна произведению числа циклов (h/d), в течение которых элементарные жилки внедрялись в первый горизонт на их среднюю толщину ($\bar{c}$). Таким образом, имеем

$$m = \frac{\hat{h}.\bar{c}}{d} \qquad (2)$$

где $\bar{c} = {}^1/_2\, c$ так как приближенно можно считать, что зияющая трещина имеет треугольное сечение и ширина ее изменяется от величины c у поверхности до 0 на глубине h. $\hat{h}$ — равно средней глубине трещин.

Из формулы (1) следует, что

$$h \approx \Delta z \approx \frac{c.G}{\tau_b} \qquad (3)$$

Таким образом, *предельная толщина ледяной жилы (m) равна произведению средней глубины растрескивания ($\hat{h}$) на среднюю ширину трещины ($\bar{c}$), деленному на среднюю скорость накопления осадков за цикл ($\bar{d}$).*

Из формулы (2) можно сделать следующие выводы :

1. Если скорость накопления осадков уменьшается, толщина жилы должна увеличиваться и наоборот.
2. Глубина растрескивания (h) и ширина трещин (c) зависят от распределения температуры в породе и от ее механических и физических свойств, поэтому при изменении последних должны меняться и поперечные сечения жил.

Следует иметь в виду, что накопление осадков между двумя последовательными растрескиваниями (d) может быть относительно неодинаково для жил, формируемых трещинами различных порядков, и увеличивается при повышении порядка трещин.

§ 5. Об определении возраста жильных льдов

Площадь поперечного сечения ледяной жилы (S) равна сумме площадей поперечных сечений элементарных жилок ($\bar{s}$), поэтому деля S на среднюю площадь поперечного сечения элементарной жилки s мы получаем число элементарных циклов развития жилок (n). Таким образом,

$$n = \frac{S}{\bar{s}} \qquad (4)$$

где

$$\bar{s} \approx \hat{h} \,.\, \bar{c} \qquad (5)$$

Среднюю ширину элементарных жилок $\bar{c}$ можно определить, подсчитывая число вертикальных полосок, образуемых пузырьками воздуха и частицами породы, собирающимися в осевой части жилки при замерзании воды в трещине. Глубину трещин ($\hat{h}$) можно оценить по формуле (3), или по температурным кривым (Б. Н. Достовалов, 1952).

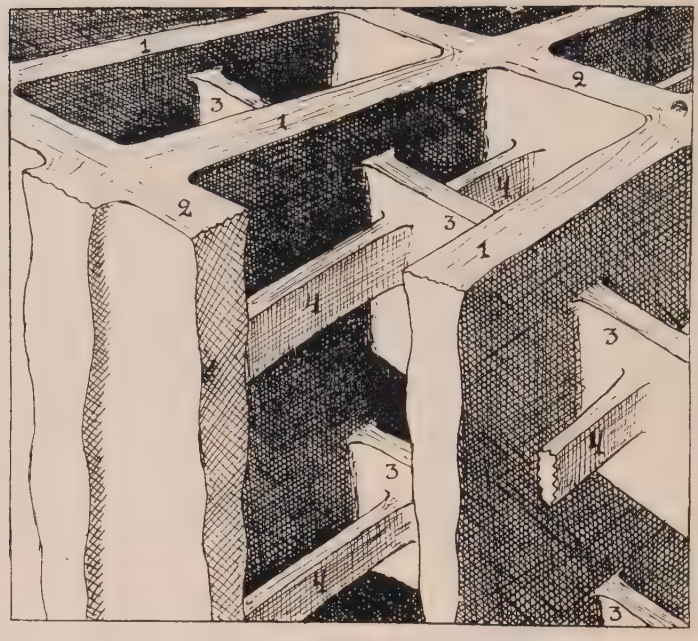

1
2
3
4

Рис. 5

9 Весенний уровень реки (Июнь)
10 Средний уровень реки (Август)
11 Нижний уровень реки (Октябрь)
12 Жильные льды
Вертикальный масштаб
0 10 20 30м

Рис. 6

В обнажении Мус-Хая на р. Яне (рис. 1) вертикальный размер крупных жил около 30 м при средней поперечной мощности 6 м. Пусть средняя ширина элементарной трещины $\bar{c} = 0{,}005$ м и ее глубина $\bar{h} = 3$ м, тогда по формуле (4) имеем

$$n = \frac{S}{\bar{h} \cdot \bar{c}} = \frac{30 \cdot 6}{0{,}005 \cdot 3} = 12000 \text{ циклов}$$

Таким образом, такие жилы росли не менее 12000 лет.

Абсолютный возраст жил равен сумме периода их роста и времени пребывания их в ископаемом состоянии и может быть во много раз длительнее периода их роста. Важно иметь в виду, что длительность роста системы жильных льдов следует определять по наиболее мощным жилам.

§ 6. Изменения роста ледяных жил в зависимости от изменения фациальных условий

В течение длительных промежутков времени сингенетического роста ледяных жил в зависимости от фациальных условий могут изменяться состав и физические свойства (τ_b, G, α) накапливающейся мерзлой породы, а также градиенты температуры ($\Delta t/\Delta z$) и скорость накопления осадков (d). В связи с этим должны изменяться как вероятность появления трещин более высоких порядков, так и эффект расклинивания или напряжения сжатия в породе.

Рассмотрение влияния вариаций всех этих условий на рост и изменение формы жил приводит к следующим выводам :

1. Изменение состава и физических свойств вмещающей породы приводит к изменению размеров отдельностей, образующихся при растрескивании. Такой процесс может искривлять вертикальные оси растущих жил, а при резких изменениях и совсем прекращать их развитие, так как новая система трещин не будет совпадать со старой.

2. При сингенетическом росте жил и возрастании градиентов температуры сетка трещин должна становиться все более мелкой в силу образования трещин все более высоких порядков, причем последние должны появляться последовательно на все более высоких уровнях по вертикали. При уменьшении градиентов процесс должен идти в обратном направлении и сначала должны исчезать трещины самого высокого порядка, а затем постепенно трещины более низких порядков. Процесс роста и форма жил в этом случае показаны на схематическом рис. 5, составленном, предполагая, что во время сингенетического роста жил произошло два похолодания на фоне относительно более мягкого климата, при котором образовывались только трещины первого и второго порядка.

Прослеживая на рисунке 5 рост льдов снизу вверх, мы видим, что в нижней части структуры при относительно мягком климате росли толко жилы 1 и 2. При увеличении минимумов и градиентов температуры в наслаивающейся породе начали образовываться трещины третьего порядка и нижняя часть нижней жилы 3. При первом максимальном похолодании к росту нижних жил 1, 2, 3 добавляется и рост жилы 4. Далее при потеплении зим процесс идет в обратном порядке и последовательно исчезают нижние жилы 4 и 3 и продолжают расти только жилы 1, 2. При новом похолодании, сменяемом затем потеплением, последовательно образуются и исчезают верхние жилы 3 и 4.

3. *Изменения скорости накопления осадков изменяют толщину жил* (формула 2) *и напряжения сжатия в растрескивающемся массиве.* При уменьшении скорости накопления осадков толщина основных жил, формируемых более вероятными трещинами первого и второго порядка увеличивается особенно сильно. Напряжения сжатия возрастают и это затрудняет появление трещин более высоких порядков и тем сильнее, чем выше порядок трещин, так как градиент температуры сначала должен снять напряжения сжатия и только уже после этого может привести к разрывающим усилиям.

Таким образом, *уменьшение скорости накопления осадков и увеличение градиента температуры действуют на появление трещин и на рост жил противоположно друг другу.* Для роста жил особенно благоприятным сочетанием является быстрое накопление осадков при высоких градиентах; наоборот уменьшение градиентов температуры и скорости накопления осадков приводят к прекращению роста жил.

Разберем с точки зрения указанных закономерностей часть обнажения жильных льдов Мус-Хая на р. Яне (рис. 1), по порядку жил соответствующую разрезу по линии АВ схематического рисунка 3.

Рассматривая подробности формы жил, мы видим, что утолщениям нижних частей первой, третьей и четвертой (слева направо) жил соответствуют более тонкие части второй и пятой жил. В средних частях первой, третьей и четвертой жил наблюдаются утоньшения, тогда как средние части второй и пятой жил утолщаются и, наконец, в верхних частях все жилы утолщаются и оканчиваются на одном уровне.

На основании указанных выше закономерностей эти соотношения толщин жил объясняются следующим образом.

Первая и четвертая жилы, формировавшиеся трещинами второго порядка, и третья, формировавшаяся трещинами четвертого порядка, являются основными и их толщина определяла режим напряжений сжатия в породе. Вторая жила формировалась трещинами пятого порядка, а пятая жила — трещинами седьмого порядка, что объясняет разницу в их толщине. Так как разница в толщине жил формируемых трещинами разных порядков, вызывается колебаниями градиентов из года в год, то климат в начальный период роста жил был неустойчивым. В то же время в этот период скорость накопления осадков уменьшалась, вызывая утолщение основных жил, увеличивая напряжения сжатия и затрудняя появление трещин более высоких порядков, поэтому утолщениям в нижней части основных жил соответствует тонкая часть второй жилы и почти полное отсутствие пятой жилы.

В период роста средних частей жил скорость накопления осадков увеличилась, основные жилы стали тоньше, напряжения сжатия уменьшились и при тех же градиентах возрасла возможность растрескивания. Поэтому средним более тонким частям второй, третьей и четвертой жил соответствуют более толстые средние части второй и пятой жил.

При росте верхних частей жил скорости накопления осадков уменьшались, вызывая утолщение основных жил. Вторая и пятая жила должны были бы при этом утоньшаться, как в нижней части, но произошло резкое увеличение градиентов (увеличение континентальности или похолодание), которое скомпенсировало влияние уменьшения скорости осадконакопления, усилило растрескивание и обусловило возможность роста и даже утолщения второй и пятой жил. Расчет времени роста верхних частей жил (§ 5) приводит к выводу, что этот период резкого похолодания продолжался не менее пяти тысяч лет.

Затем градиенты уменьшились (потепление или уменьшение континентальности), растрескивание ослабело, верхняя поверхность жил оторвалась от нижней поверхности слоя летнего протаивания и они перешли в ископаемое состояние.

Таким образом, предложенная физико-механическая теория формирования систем ледяных жил позволяет объяснить сложные детали разрезов, сделать выводы о смене фациальных условий и дать некоторые оценки возраста жильных льдов.

В качестве второго примера рассмотрим схему рельефа местности в низовье р. Индигирки (рис. 6). Изложенные выше закономерности развития систем ледяных жил позволяют сделать следующие выводы об истории формирования этого разреза.

В прошлом пойма находилась на современном уровне верхних частей древних ледяных жил второй террасы и на ней отлагались пески, подстилающие в настоящее время эти льды. Климат был суровым; эти пески были мерзлыми, но льды не формировались вследствие неблагоприятного состава породы. Затем скорость течения реки уменьшимась, на слое песков начал накапливаться слой суглинка и в нем начали формироваться нижние части ледяных жил второй террасы. Этот процесс протекал при относительном равновесии между скоростью накопления осадков и скоростью *опускания местности* в течение 10-20 тысяч лет.

Когда опускание прекратилось, поверхность развития льдов стала выходить в стадию первой террасы, рост льдов прекратился, как описано выше для обнажения Мус-Хая (рис. 1).

Следующим этапом развития явилось опускание базиса эрозии, врез русла реки на глубину около 40 м и превращение горизонта с древними ледяными жилами в относительно высокую вторую террасу. Затем процесс сингенетического роста ледяных жил повторился в меньшем масштабе и образовались

породы и льды первой террасы, при этом условия сингенеза сохранялись в течение 4-5 тысяч лет.

Размеры полигонов на первой террасе крупнее, чем на второй, поэтому можно заключить, что она формировалась при более мягком климате.

Затем произошел новый врез реки метров на 20 и после него начали накапливаться отложения и льды уже современной поймы, на некоторых участках которой они растут и в настоящее время уже в течение по крайней мере 1500 лет.

Суммируя длительности аккумуляции террас (20000 + 5000 + 1500 лет) и считая, что на врезы потребовались такое же время, мы получаем нижний предел абсолютного возраста льдов второй террасы около 50000 лет, но вероятность отсутствия тектонических движений в течение неопределенных промежутков времени может повысить их возраст в несколько раз.

Подводя итог, мы видим, что образование крупных жил на нескольких различных уровнях необходимо связано с чередующимися и в данном случае затухающими опусканиями и поднятиями территории; размеры жил позволяют оценить длительность аккумуляции террас; изменения числа порядков трещин и рост жил на разных горизонтах одной и той же террасы свидетельствуют об изменении климатических условий. Существование древних льдов с абсолютным возрастом в несколько десятков тысячелетий исключает на обширных территориях распространение покровного оледенения. Дальнейшее более детальное изучение форм и структур жильных льдов несомненно увеличит число подобных выводов.

VARIABILITY OF PHYSICAL CHARACTERISTICS OF SNOW COVER ACROSS CANADA

GAYNOR P. WILLIAMS

Abstract

The results from a survey of the physical characteristics of snow cover across Canada have been summarized. Regional variation of snow density, snow hardness, and snow grain-size are shown graphically, as well as the probability values of extremes for these snow characteristics.

At Oslo in 1948 Klein ([1]) presented the preliminary results of a survey of the physical characteristics of snow cover across Canada to the International Association of Scientific Hydrology of the International Union of Geodesy and Geophysics. D.C. Pearce and L.W. Gold ([2]) extended the summary made by Klein for presentation at the I.U.G.G. meeting in Brussells in 1951. The purpose of this present analysis is to summarize additional results from this snow survey and to show the variability that can be expected in snow-cover properties across Canada.

The original objective of the survey was to carry out periodic measurements of the physical characteristics of the snow cover at a number of observation stations to obtain useful data applicable to a wide variety of winter problems. The survey, sponsored by the Associate Committee on Soil and Snow Mechanics of the National Research Council of Canada, has been carried out with the co-operation of the Meteorological Division, Department of Transport and various pulp and paper, and power companies across Canada.

The instruments used to measure the characteristics of the snow during this program developed into what is now known as the «Canadian Snow Kit». A detailed description of this kit, including the methods used to measure snow characteristics is contained in papers by Klein and Pearce and Gold ([1], [2]).

Observation Stations

Figure 1 shows the location of the observation stations providing the data which are now analysed. Snow observations have been made at several other stations which were not included because the data were often insufficient for detailed analysis.

The survey was limited to two types of stations: exposed stations with a flat, unprotected test area free from disturbances due to trees, buildings etc., and sheltered stations with a test area sheltered from wind action. At each station a suitable area of about 500 square feet was selected for the semi-monthly observations of the snow profile.

It is considered that the stations shown in Fig. 1 cover a wide enough range in climatic conditions to be representative of snow characteristics likely to be encountered in Canada.

Data Used in Analysis

The snow characteristics analysed were snow density, hardness, and grain size. Snow density gives a measure of the relative proportion of ice and air in snow.

Fig. 1 — Location of the observation stations across Canada.

It was measured by inserting a 250-CC sampler horizontally into the snow-cover profile and weighing the sample of snow obtained. Though density measurements were obtained from the ground to the snow surface at regular, layer intervals, only density measurements from 6 to 18 cm below the snow surface were used in this analysis. Choosing a specified position in the snow cover can be compared to the standardization of the height of meteorological instruments so that readings from different stations can be compared.

Snow hardness as measured wiht the NRC hardness gauge is not easy to define. However, by measuring the stress under a compressive load at which initial collapse of the snow structure occurs, some measure of the strength of deposited snow can be obtained ([3]). The measurement is obtained by pushing a disc horizontally against the exposed snow layer and measuring by means of a graduated spring the force in grams per sq cm at which collapse takes place. Only hardness measurements from 6 to 18 cm below the snow surface have been used in this analysis.

Grain size is estimated by comparing the size of the grains with the spacing of lines on a plate ruled in millimetres. As there were insufficient grain-size observations in any specified layer, grain-size data were taken from all available layers.

To describe a snow cover properly, measurements on grain shape, free water content and snow temperature are also needed, but these were not included in this analysis. Measurements of snow temperature obtained semi-monthly on snow profiles at different stations do not give much information for statistical analysis. Because of the difficulty of measuring free water content under field conditions, this snow characteristic was seldom measured by the snow observers. A preliminary analysis of grain-shape data indicated that there was often too much bias in the results to warrant including them in this study.

The numbers of observations are given in Table I. Over 600 density and hardness observations and over 2000 grain-size observations were used. It was considered that

there were enough observations to enable a proper analysis to be made of these three properties of snow and thus to show the regional variation that could be expected under a wide variety of climate and site conditions.

TABLE I

Analysis of snow survey data from stations

Station	Years of observations	No. of density observations*	No. of hardness observations*	No. of grainsize observations**
Sheltered				
Shawbridge	3	36	44	84
Forestville	3	37	49	152
Fredericton	3	29	28	80
Ottawa	1	39	61	49
Aklavik	1	73	72	74
Exposed				
Old Glory	3	44	48	281
Resolute	5	209	128	693
Winnipeg	3	58	60	168
Ottawa	5	33	26	123
Gander	3	31	62	74
Aklavik	3	31	60	385
Grand Total		620	638	2163

*6 to 18 cm from surface
**all layers

Method of Analysis

The assembled data were arranged in class intervals. For example, all density data were tabulated in class intervals from 0 to 0.05, 0.05 to 0.10 up to 0.45 to 0.50 gm/cm^3. Once the data were arranged in this form they were plotted on arithmetic probability paper. On this type of paper a normal distribution will appear as a straight line ([4]). This form of plot is a convenient way of comparing snow survey results from several regions on one graph. The same information could have been presented by standard distribution histograms but it would have been difficult to compare the histograms graphically on one drawing. In addition, this form of presentation gives the probability of minimum and maximum values of certain snow characteristics in various regions; information often useful in applied problems.

Density Variability

Figure 2 shows the snow density data from 6 to 18 cm below the snow surface plotted on arithmetic probability paper.

For all sheltered stations, density values were under 0.41 gm/cm³ 90 per cent of the time. The only station that showed somewhat different variability was Aklavik where densities were less than 0.28 gm/cm³ 90 per cent of the time. This station has smaller density values because frequent depth hoar formations result in a comparatively low-density snow type.

Snow density shows more variation for exposed stations. At Resolute density

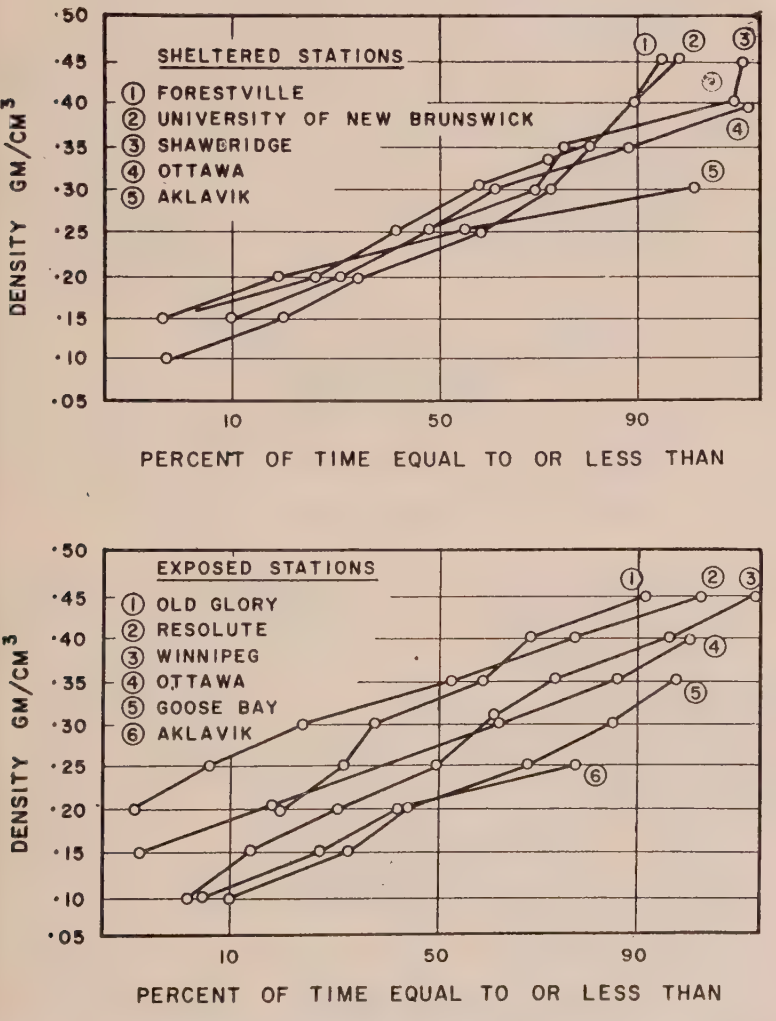


Fig. 2 — Snow density data from 6 to 18 cm below the snow surface plotted on arithmetic probability paper.

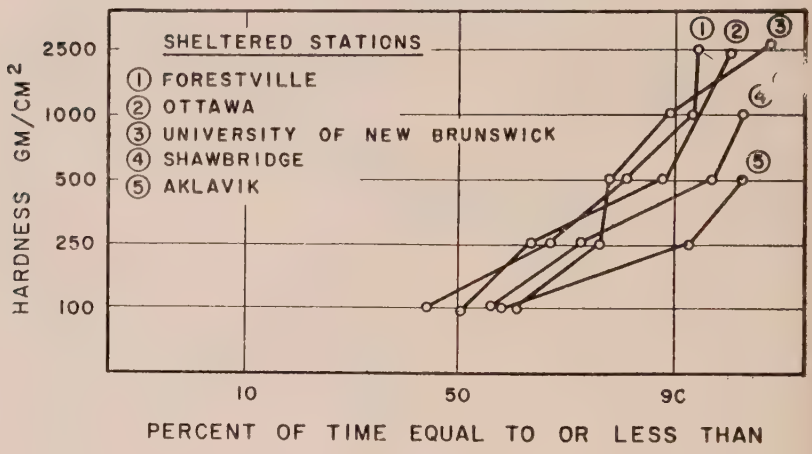
values are greater than 0.26 gm/cm³ 90 per cent of the time. Again frequent depth hoar formations at the Aklavik site result in comparatively low density snow.

Figure 2 shows that snow density in Canada generally varies from 0.05 to 0.45 gm/cm³, with values outside this range rarely reported. Varying exposure to wind action causes greater density variation at exposed than at sheltered sites.

Hardness Variation

Figure 3 shows a probability plot of hardness values in snow covers, 6 to 18. cm below the surface, at sheltered and exposed stations across Canada.

At sheltered stations hardness values are under 1000 gm/cm² 90 per cent of the time; and under 100 gm/cm² 40 per cent of the time. At Aklavik hardness values



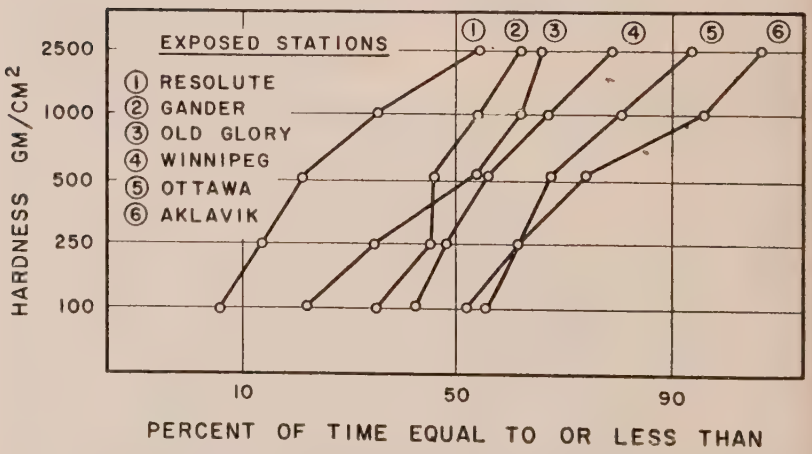


Fig. 3 — Probability plot of hardness values in snow covers, 6 to 18 cm below the surface, at sheltered and exposed stations across Canada.

are under 250 gm/cm² 90 per cent of the time. Aklavik, with a cold, stable winter, has few freeze-thaw periods during the winter months; this appears to be the main cause of higher hardness values at the other sheltered stations.

At exposed stations there is a much greater variation in hardness values. It is interesting to note that the two most northern stations report the greatest extremes in hardness values. At Resolute, hardness values of 2500 gm/cm² or greater occur at least 40 per cent of the time; at Aklavik, hardness values of 100 gm/cm² or less occur at least 55 per cent of the time.

Grain-size Variations

Figure 4 shows a probability plot of grain-size determinations from all snow layers at sheltered and exposed stations across Canada.

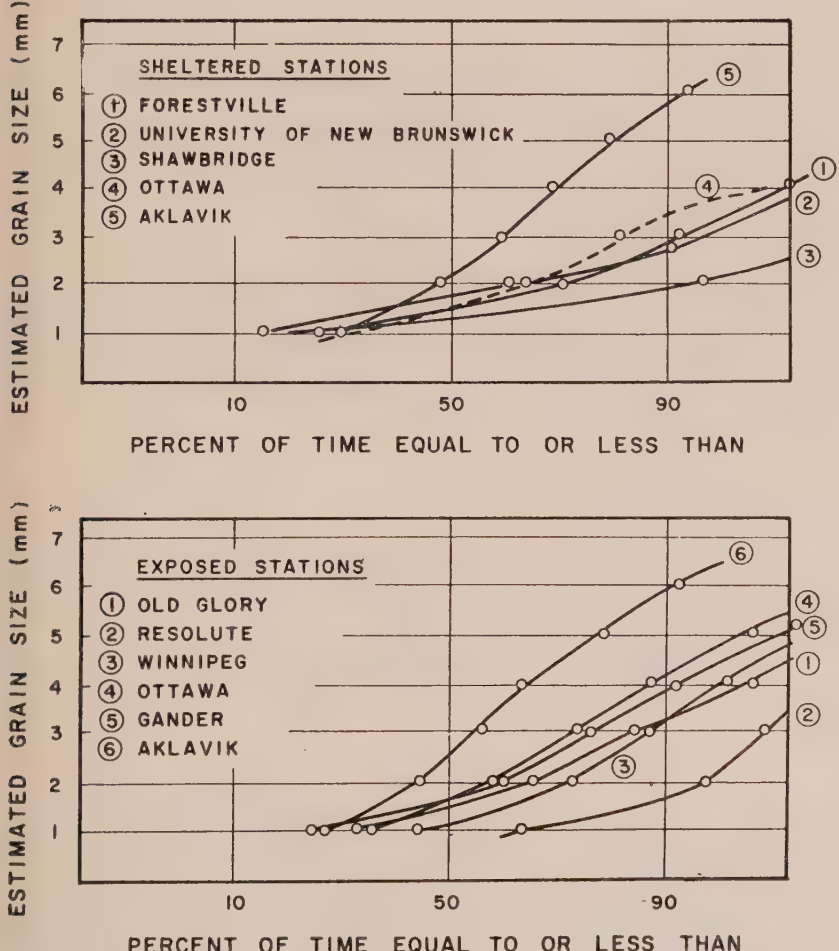


Fig. 4 — Probability plot of grain-size determinations from all snow layers at sheltered and exposed stations across Canada.

Though grain size is sometimes difficult to measure, and thus subject to personal bias, Figure 4 shows that grain-size curves are quite similar at all sheltered stations except Aklavik. At Aklavik, approximately 20 per cent of the grain sizes were reported over 5 mm. At all other sheltered stations 90 per cent of the values reported were under 3.5 mm.

The grain-size curves are almost identical for all exposed stations with the exception of Aklavik and Resolute.The grain-size curve for the exposed site was almost identical to the curve for the sheltered site at Aklavik. At Resolute 90 per cent of the observations were under 1.5 mm.

Figure 3 indicates that, except for Aklavik, 90 per cent of all snow grain sizes reported were under 4 mm. and approximately 50 per cent of all grain sizes reported were under 1.5 mm.

Discussion of Results

The analysis of snow survey results by the use of probability paper can be compared to climatological analysis. Just as a description of the climate of a region will not give much information as to meteorological conditions at a particular time or place, neither will a climatological analysis of snow survey results give accurate information about snow properties at a particular time or place. Figures 2, 3, and 4 however, do give a representative picture of the regional variability of specific snow properties such as density, hardness, and grain size.

It is considered that by giving the probability of occurrence of certain snow properties at different locations some useful information has been gathered for those engaged in snow studies. Density values provide some information from which snow thermal conductivity can be estimated. From hardness values, designers of over-snow equipment will be able to make assumptions regarding the extremes likely to be encountered. Grain-size values provide better understanding of the variability of snow as a material.

Acknowledgements

The author wishes to acknowledge the work of the observers who, over a period of years, measured the different snow properties at stations across Canada, often under adverse weather conditions. This report would not have been possible without their co-operation and effort.

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PHYSICAL PROPERTIES OF THE SNOW COVER

G. K. SULAKVELIDZE

Summary

1. The snow cover is considered to be a porous medium containing water in three conditions: solid, liquid and gaseous, with the prevalent solid phase. Physical properties of snow are considered by the liquid and gaseous contents in the snow cover and by its density.

2. Snow cover compression and extension indices are determined experimentally. Investigations show that if a pressure on the snow cover does not exceed 50 gr per cm^2, the latter experiences a plastic deformation. With a further increase of pressure the snow compression continues due to brittle fracturing of snow grains and crystals.

The report gives data on a value of the coefficient of internal friction and cohesion forces and that of snow toughness depending on its density.

3. Besides, there are data concerning the character of solar radiation reflection by the snow surface and its absorption by the snow cover, as well as values of spectral coefficients of radiation absorption and reflection from and in the snow cover.

It is found that in cases when an angle of descent of a ray with the snow cover surface is less than 70°, the indicatrix of reflection from the snow surface has no mirror component which appears in a fall plane at an angle of descent of over 70°.

4. Owing to experiments the coefficient value of the heat conduction of the snow cover and its dependence on the snow density are defined more accurately.

An equation of full heat exchange in a snow mass is drawn with allowance for the energy of phase transitions for three cases:

a) if the snow temperature is less than —25° C;

b) if the temperature varies from 0° to —25° C;

c) if highly damped snow is frozen.

5. An equation of a complete heat balance of the snow cover is drawn with allowance for the energy of phase transitions in the snow thickness depending on its density and moisture content.

6. An equilibrium equation of snow masses on mountain slopes for loose and packed snow is drawn and Balabuev-Sulakvelidze avalanche diagrams for forecasting avalanches are given.

7. A relation is determined between physical and mechanical properities of snow (internal friction angle, density, etc.), angle of slope, length of the slope and velocity of avalanche.

8. A pressure originating during the avalanche impact on motionless obstacles is calculated on the basis of physical properties and mass of the avalanche snow cover, angle and length of the slope.

9. The author gives the results of comparing the force of the avalanche impact on motionless obstacles with computation data. A nomograph for determining the force of the avalanche impact on motionless obstacles is given.

СНЕГ И ЕГО СВОЙСТВА

Г. К. Сулакшелидзе.

Снежный покров, несмотря на то, что первые опыты по исследованию его физических свойств насчитывают столетнюю давность (30, 34, 37), изучен совершенно недостаточно.

Задача, которую мы ставим в этой работе заключается в том, чтобы найти значение коэффициентов, характеризующих физические свойства снега как функцию его плотности и влажности.

СЖАТИЕ СНЕГА

В предыдущих исследованиях (7, 8) более и менее подробно были изучены механические свойства снега (сжатие и растяжение, сопротивление разрыву). Однако количественная связь между первоначальной плотностью снега, приложенной нагрузкой, временем и величиной деформации были получены лишь для нагрузок порядка 80-200 г/см², которые обычно возникают при искусственном уплотнении снега (9).

При нагрузках до 45 г/см² уплотнение снега происходит за счет более компактного распределения снежинок и вытеснения находящегося между ними воздуха. При этом возникает пластическое деформация и перераспределение снежинок за счет преодоления внутреннего трения.

Для установления зависимости между увеличением плотности снега и приложенной к нему нагрузки в термокамере, где можно было поддержать постоянную температуру в интервале от 0 до —15° С, к пяти образцам снега одной и той же плотности прикладывались нагрузки от 5 до 50 г/см² и исследовалось относительное укорачивание образца снега.

Из этих опытов было получено, что относительное увеличение плотности снега при нагрузках не свыше 30 г/см² и температурах от 1° до —15° может быть представлена в виде:

$$\frac{\Delta\rho}{\rho} = \left[(0{,}38 - 1{,}88\rho_o)\sigma + (11{,}8 - 3{,}88\rho_o)\right]\cdot 10^{-6}\cdot t + (1{,}5 - 3{,}6\rho_o)\sigma\cdot 10^{-2} + (0{,}16 - 0{,}78\rho_o)\cdot 10^{-2} \qquad (1)$$

где t — время в минутах, отсчитываемое от момента приложения нагрузки, σ — напряжение, $\Delta\rho$ — приращение плотности, ρ_o — первоначальная плотность снега. Приведенную формулу можно применить после двух часов со времени приложения нагрузки.

При внезапных значительных нагрузках, возникающих при ударе снежной лавины о неподвижное препятствие, деформация сжатия снега приходит, главным образом, за счет хрупкого излома, мгновенной режеляции. На основании экспериментальных исследований, проведенных в 1949 — 52 г.г. была получена зависимость между конечной плотностью снега и величиной нагрузки при значениях последней, превыщающей 0,5 кг/см².

$$\Delta\rho = (3{,}10 - 7{,}81\rho_o)\rho_o P \qquad (2)$$

где ρ_o — первоначальная плотность снега, $\Delta\rho$ — приращение плотности, а P — давление в кг/см². Формула (2) может быть применена для сухого снега при начальных плотностях от 0,11 до 0,37 г/см³. Средняя погрешность ь определении величины

$$\frac{\Delta\rho}{\rho_o}$$

не превышает 2 % :

Уравнение (1) может быть применено при расчете оседания снега под действием собственного веса, что для плотности снега на глубине Z приводит к выражению вида:

$$\rho_{(z)} = \rho_o\left[(0{,}38 - 1{,}88\rho_o)\cdot 10^{-6}\, t + (1{,}5 - 3{,}6\rho_o)\,\sigma\, 10^{-2}\right]Z \qquad (3)$$

В отличие от соответствующих выражений, полученных Абэ, Шепелевским (1908, 1937 г.г.) и др., уравнение (3) учитывает изменению плотности со временем (см. Вейнберга).

Для определения плотности снега формула (3) не может применяться в районах с интенсивной радиацией, способствующей перекристаллизационным и режеляционным процессам и образованию « ветвистой » структуры режелированного снега внутри покрова, чем в основном и определяется плотность снега на этих участках (35, 40).

РАСТЯЖЕНИЕ СНЕГА

В результате опытов, проведенных, как в природных, так и в лабораторных условиях, было установлено, что величина деформации снега при растяжении не может быть представлена как однозначная функция его состояния (плотности, влажности и температуры).

Образцы сухого снега, выпадавшего в штилевую погоду, подвергались в лаборатории колебаниям температур от —3° до —10°, и даже на 6-е сутки после взятия проб их сопротивления разрыву равнялось нулю.

Снежный покров со смерзшимися или же режелированными зернами в некоторых случаях выдерживал разрывающее напряжение в 200 г/см² в течение 2 суток. Когда снежный покров оказывал хотя бы небольшое сопротивление разрыву, при детальном исследовании обнаруживались системы сросшихся, режелированных, смерзшихся снежинок.

Таким образом, при растяжении снега мы имеем дело с деформацией системы ледяных ветвей, а не снежных кристаллов.

Величина разрывающего напряжения σ_e зависит, главным образом, от физико-механических и тепловых процессов в снежном покрове, предшествовавших опыту. Таким образом, деформация снега при растяжении не является однозначной функцией его температуры и плотности, а зависит от погодных условий.

Величина относительной деформации снега при растяжении, даже для фирнизированного наста, не превышала 70 %.

ВЯЗКОСТЬ СНЕГА

Вязкость снега нами определялась двумя методами. Первый-метод течения снега под действием внешней нагрузки, описанный Гоффом и Оттеном 1932 г., приводит к быстрому уплотнению снежного покрова (8, 9), в результате чего определяется коэффициент взякости фирна, а не снега.

Второй метод, заключался в изучении микросползания снега под действием собственного веса и вычислении коэффицента вязкости по йзвестной формуле:

$$M = \frac{g\rho \sin \alpha \left(\frac{h}{2}\right)^2}{\Delta V_z} \qquad (4)$$

где M — коэффициент вязкости, $g\rho$ — удельный вес, h — глубина снега, α — уклон наклона склона, ΔV_z — изменение величины скорости микросползания с глубиной. Скорость микросползания определялась путем измерения перемещения оси скважин, заданных в снежном покрове на склоне с уклоном 30°—48°, внутрь которых засыпалась краска. На 4-12 сутки после закладки скважины раскапывались и измерялись перемещение оси скважины, и распространение краски в толще снега, определяющее диффузионные процессы.

Зависимость M от ρ, приведенная в табл. I, может быть представлена в виде следующего выражения:

$$M = 2{,}5 \cdot 10^4 \, l^{27{,}65\rho} \qquad (5)$$

Табл. 1.

ρ 2/см³	M $\frac{\text{дина сек}}{\text{см}^2}$	Примечание
0,15	1,0 . 10	
0,20	2,8 . 10^9	
0,30	3,4 . 10^{10}	
0,35	1,0 . 10^{13}	
0,50	1,3 . 10^{13}	по данным акад. Завриева 5,29
0,90	1,2 . 10^{14}	по данным Дили (13)

Коэффициент трения, разрывающее напряжение и сила сцепления в снежном покрове.

Все эти свойства были подробно изучены в 1928-1937 г.г. А. Г. Гоффом и и Г. Ф. Оттеном (8, 9), что позволяет нам коснуться лишь некоторых особенностей, которые в свое время не были отмечены упомянутыми авторами.

Во всех наших опытах трение покоя на 5 % и больше превышает трение движения, что указывает на невозможность представления снежного покрова как сыпучей среды.

Для свежего сухого снега разрывающее напряжение близко к нулю.

Для поверхностного слоя (наста) величина разрывающего напряжения достигает 200 гр/см².

Поверхностный слой создается на второй — третий день после снегопада, но к этому времени у поверхности почвы (или погребного наста) уже образуется глубинный иней, способствующий срыву лавин (34, 38, 39, 40).

Поэтому в ряде работ указывается на лавинную опасность при подрезании поверхностного пласта лыжней (см., например, Зелигмана).

В 1939 году было отмечено, что внезапные небольшие потепления на Большом Кавказе приводят к резкому увеличению схода лавин (25). Одно-двухдневное потепление в феврале месяце, конечно, не может привести к увлажнению всей толщи покрова на Большом Кавказе, а вызывает небольшое таяние лишь поверхностного слоя, приводящее к нарушению связей между отдельными кристаллами и резкому уменьшению σ_c наста, что в свою очередь способствует срыву лавин.

Ниже приводится сводная таблица величин, характеризующих механические свойства снежного покрова (8, 9).

Табл. 2.

Характер снега	Угол внутреннего трения s	коэффиц. трения f	разрыв напр. σ_c		
			мним	максим	среднее
1. Сухой снег.					
1. Глубинный иней	5°	—	0	10	10
2. Свежий снег	10°-20°	0,-0,10	10	20	10
3. Уплотнившийся снег	30°	0,40-0,50	75	500	300
4. Фирн	50°-60°	—	—	—	—
2. Мокрый снег.					
1. Свежий снег	20°-30°	0,13-0,35	100	300	250
2. Уплотнившийся снег	40°	0,30-0,48	300	1000	500
3. Фирн	60°-70°	—	500	5000	900

АЛЬБЕДО СНЕГА

Альбедо снежного покрова изучалось нами с помощью пиранометров Ю. Д. Янишевского, фотоэлементов и желатиновых фильтров с избирательным пропусканием. Установлено, что сухой снег различных видов, при прочих равных условиях, имеет одинаковое альбедо прямой радиации в пределах точности измерения — 2 %

Альбедо рассеянной радиации сухого снега зависит от характера облачности и изменяется от 60 % при высокой облачности до 80 % в случае приземного тумана. Для влажного и загрязненного свежего снега величина альбедо изменяется от 45 % (сильно загрязненный снег) до 60-65 % (слегка увлажненный свежий снег). В случае прямой радиации альбедо снега зависит от угла падения луча. В результате исследований (33) были получены данные, приведенные, в таблице 3. Каждое приведенное значение является средним из 50 наблюдений.

Таблица 3

Угол падения	0°	10°	20°	30°	40°	50°	60°	70°	80°	90°
Отношение отраженной радиации в % к падающ. $\frac{I_o}{S_n} \cdot 100 = A$	59,1	59,1	159,4	60,0	63,0	66,0	69,8	74,5	82,5	97,2

По данным этой таблицы получена зависимость величины отраженной радиации от угла подения луча:

$$I_o = 95{,}5 - 69{,}8 \cos \alpha\ 33{,}8 \cos^2 \alpha, \qquad (6)$$

где I_o — отраженная радиация в % к падающей, α — угол падения прямой радиации.

Формула (6) об'ясняет расхождение в величинах альбедо, полученных Н. Н. Калитиным, Н. Т. Черниговским. Данные обоих авторов хорошо совпадают со значением, вычисленным по формуле (6), если учесть разницу в высотах Солнца на Эльбрусе и на острове Уединения, где проводились эти исследования (11, 12, 41).

Аналогичная работа была проведена по определению величины отражения солнечной радиации для шести участков спектра (см. табл. 4).

Таблица 4.

Угол падения / данны волны	80°	75°	70°	60°	50°	45°	30°	15°	10°	0°
1000	55%	49%	—	40%	—	32%	24%	23%	22%	22%
795	71	62	58	54	52	50	48	45	44	44
590	71	62	—	54	—	51	50	49	48	48
560	73	70	—	62	—	55	52	50	50	50
500	88	78	74	70	—	66	62	62	62	62
425	82	72	63	56	—	52	49	44	44	44

П. П. Кузмин и др. (15, 16) отрицают зависимость между углом падения солнечной радиации и величиной альбедо. Н. М. Писякова в 1947 г. получила зависимость между альбедо и высотой солнца (26).

ИНДИКАТРИССА ЯРКОСТИ СНЕЖНОГО ПОКРОВА

Индикатрисса яркости снежного покрова была определена Криновым в 1951 г. Однако эти определения требовали дальнейшего уточнения (16, 17), в 1950-1953 г.г. нами были проведены работы по определению индикатриссы

яркости снежного покрова для различных углов падения солнечного луча. Яркость определялась вдоль следа вертикали солнца на снежном покрове и вдоль прямых составляющих со следом вертикала углы в 45° и 90°. Яркость покрова в этих направлениях определялась при высотах солнца в 15°, 30° и 60°.

Результаты сведены в таблице 5.

Таблица 5.

Угол над горизонтом, вдоль которого определялось яркость	Отношение яркости снегак максимальной яркости в %			Высота солнца
	15°	45°	60°	
15°	35	80	88	Отсчеты углов вдоль которых определялось отражение; велись со стороны солнца
30°	35	84	91	
45°	32	81	86	
60°	35	83	86	
60°	35	83	86	
90°	38	91	97	
105°	40	91	99	
120°	46	91	99	
135°	53	95	99	
150°	64	99	100	
165°	100	100	88	

Анализ таблицы приводит к следующим выводам: а) в вертикале солнца, при высоте последнего в 15° и меньше, снежный покров проявляет свойства зеркального отражения (класс Б по Шаронову (48), б) в вертикале сонца, при высоте последнего больше 15°, снег обладает матовым отражением (класс А по Шаронову), в) во всех остальных направлениях снег характеризуется также матовым отражением независимо от высоты солнца.

Данные Мидля и Монгола совпадают с приведенными результатами. Опыты Зауберера (50) также подтверждают наличие зеркального отражения у снежного покрова при больших углах падения, хотя Тамс отрицает этот эффект (54).

В 1954 году. Были сняты индикатриссы яркости снега для длин волн 619*мм* и 496*мм* при высоте солнца в 15° и 18°, которые подтвердили характер зеркального отражения снега при больших углах падения солнечного луча (2).

ПОГЛОЩЕНИЕ СОЛНЕЧНОЙ РАДИАЦИИ В ТОЛЩЕ СНЕЖНОГО ПОКРОВА

Исследования поглощения солнечной радиации в толще снежного покрова проводились аппаратурой, основанной на принципе термопиранометров (33, 35).

Результаты исследования коэффициента поглощения К интегральной радиации в зависимости от плотности ρ снежного покрова сведены в таблице 6.

Таблица 6.

ρ г/см³	0,10	0,19	0,39	0,59	0,65
К см²/г	1,19	1,03	0,86	0,74	0,72

Поглощение радиации в толще снежного покрова происходит по экспоненциальному закону, причем интенсивность радиации I_z на глубине Z можно представить в следующем виде:

$$I(\rho Z) = (I_n - I_o)l^{-K\rho Z} \qquad (7)$$

По данным таблицы (6) получаем следующую зависимость между К и ρ:

$$K = 1{,}31 - 1{,}11\rho$$

Подставив выражение для К в формулу (7), получим:

$$I(Z\rho) = (I_n - I_o)l^{-(1{,}31 - 1{,}11\rho)\rho Z} \qquad (8)$$

Значения К, полученные Н. Н. Калитиным, П. П. Кузьминым, Тамсом и Зауберером (1936-1948 г.г.), укладываются в формулу (11, 19, 20, 50, 54).

Результаты определения коэффициента поглощения для различных участков спектра в зависимости от плотности снега приведены в таблице 7.

Таблица 7.

	0,10	0,19	0,39	0,59	0,65	средняя величина
725	1,20	1,11	0,94	0,79	0,79	0,97
590	1,34	1,22	1,01	0,80	0,80	1,07
560	2,00	1,73	1,23	0,82	0,82	1,39
500	1,04	1,00	0,80	0,76	0,72	0,90
425	1,70	1,50	1,16	0,81	0,80	1,25

КОЭФФИЦИЕНТ ТЕПЛОПРОВОДНОСТИ СНЕЖНОГО ПОКРОВА

Зависимость между коэффициентом теплопроводности и плотностью снега была впервые определена Абельсом Г. Ф. (1) методом затухания температурных волн в толще покрова (1893). Однако, работа Окада, Янсена, Хильстрома (1902-1904, 1914), Кузьмина А. А. и Кондратьевой Л. С. (1939) показали, что исследования Абельса требуют уточнения (3, 14, 18, 47).

Анализ условий опыта Абельса показал, что средняя погрушность в этих исследованиях была не меньше 16 % от измеряемой величины.

В 1955-1956 г.г. Иосида (46), предложил новый метод определения теплопроводности снега. Металлический сосуд, заполненный снегом, термостатировался при температуре u_1, герметизировался и соединялся с микроманометром. В дальнейшем сосуд опускался в раствор с постоянной температурой u_2. Время наступления изотермии в снеге при температуре u_2 определялось давлением воздуха в сосуде микроманометром.

При разности температуры снега $(u_1 - u_2)$, времени, необходимом для изменения температуры снега, а также по форме сосуда, принимая что теплопроводность сосуда бесконечна велика по сравнению с теплопроводностью снега, Иосида вычислил величину коэффициента теплопроводности снега. По измерениям Иосида коэффициент теплопроводности оказался равным $1{,}37.10^{-4}$.

Преимущество метода Иосида перед другими методами заключается в том, что для определения температуры снега не приходится вносить в последний измерительные приборы, искажающие картину теплообмена. Однако метод Иосида требует двух поправок, которые в работе не оговорены. Водяной пар в толще снежного покрова находится все время в состоянии насыщения, определяемого температурой снега. Изменение температуры в опытах Иосида вызывает конденсацию пара воды. Конденсация водяного пара, с одной стороны, приведет к уменьшению парциального давления (в опытах Иосида при перепаде температур около 6° вблизи 0° изменение давления водяного пара должно было составлять около 2 мб). С другой стороны, конденсация водяного пара вызовет дополнительный нагрев снега, который в опытах Иосида со-

ставляет около 0,03-0,05° С. Оба эти явления обуславливают замедление уменьшения давления смеси воздуха и пара. Возможно, что внесение этих поправок значительно увеличит коэффициент теплопроводности, полученный Иосида, который на порядок меньше данных Абельса, Янсена, Кузьмина и др. авторов.

Как следует из приведенных работ, центральной задачей в области исследования теплового режима снежного покрова было принято считать задачу нахождения коэффициента теплопроводности снежного покрова в зависимости от его структурных особенностей.

В настоящее время (23, 32, 35, 46) выяснено, что процесс теплопередачи в толще снежного покрова представляет собой весьма сложную картину и не может быть описан каким-либо одним механизмом теплопроводности. Представление о справедливости уравнения молекулярной теплопроводности Фурье для снежного покрова оправдывается не во всех случаях. Важное значение в процессе теплообмена имеют также явления термической диффузии, поглощения радиации и фазовых превращений. В зависимости от внешних условий, эти процессы приобретают различное значение. Анализ общего уравнения теплопроводности (23, 35), а также значительного количества экспериментальных данных указывает на целесообразность раздельного рассмотрения частных случаев теплопередачи в снежном покрове.

а) Сухой снег при температуре ниже 15°С.

Ввиду небольшого количества паров, содержащихся в снежном покрове, в данном случае тепловым эффектом фазовых превращений можно пренебречь. Процесс теплопередачи при этих условиях может быть описан уравнением молекулярной теплопроводности. При этом естественно ожидать наличия простой связи между коэффициентами теплопроводности и плотностью снега, как и для большинства других пористых материалов. Задача исследования теплопроводности снега в данном случае значительно упрощается и сводится к отысканию зависимости коэффициента теплопроводности от плотности снега.

Наиболее подходящим методом для установления этой зависимости признаны методы, основанные на анализе нестационарного температурного режима. Нами был применен метод мгновенных источников тепла. Анализ этого метода дан в работе (22, 35).

Успех эксперимента зависит, в основном, от правильного выбора способа регистрации температурного максимума и точности измерения расстояния от источника до регистратора. При плотностях сухого снега от 0,10 до 0,35 коэффициент теплопроводности в системе CGS имеет следующее значение:

$$K = 1.22 \cdot 10^{-3} \rho \qquad (7)$$

б) Сухой снег при температуре от 0° до — 15° С.

В этом случае в переносе тепла участвует также диффундирующий водяной пар, и появляются источники и стоки тепла в результате фазовых превращений. Если допустить, что места фазовых преобразований непрерывно распределены в толще покрова и принять их за источники тепла, то общее уравнение теплопроводности снега (32, 33) принимает следующий вид:

$$\frac{\partial u}{\partial t} = \left[\frac{\lambda c + \gamma D f'(u)}{1 + \gamma f'(u)}\right] \frac{\partial^2 u}{\partial z^2} + \gamma f''(u) \left(\frac{\partial u}{\partial x}\right)^2 \qquad (8)$$

где λ_c — температуропроводность сухого снега, D — коэффициент диффузии водяного пара в покрове, $f(u)$ — количество водяного пара в единице об'ема снега, $\gamma = \frac{R - L}{c\rho}$, R и L скрытая теплота испарения и плавления соответственно, $c\rho$ — об'емная теплоемкость снега.

Таким образом, исследование теплообмена в толще снежного покрова в данных условиях, наряду со знанием λ_c требует дополнительных сведений об упругости водяного пара в толще снега и о величине термодиффузии.

Отсутствие достаточных экспериментальных данных не позволяет применять уравнение для исчерпывающего решения практических задач. Однако, неоднократные указания на интенсивную перекристаллизацию снежного покрова (21, 28, 40, 46) дают основание предполагать, что применение уравнения внесет значительные коррективы в наши представления о теплопередаче в снежном покрове, т.к. уже простая оценка этого эффекта с точки зрения теплового баланса указывает на большие изменения в распределении температур (35).

в) Промерзание влажного снега.

Необходимо рассмотреть два случая промерзания влажного снега: промерзание при малом влагосодержании (до 7 %) и промерзание сильно увлажненного снега.

В первом случае следует учитывать теплоту фазового перехода водяного пара диффундирующего к промерзшей части от прилагающих снежных слоев.

Во втором случае тепловым эффектом испарения и сублимации можно пренебречь по сравнению с теплотой, идущей на промерзание влажного снега. Как было показано (23), скорость промерзания и распределение температур в промерзшей зоне при этом могут быть с достаточной степенью охарактеризованы решениями задачи Стефана.

Если обозначить координату фронта промерзания через ξ а время — через t, то при постоянной отрицательной температуре на поверхности покрова связь между ξ и t выражается следующими уравнением: $\xi = \beta\sqrt{t}$

Коэффициент β зависит от состояния снега и температуры его поверхности:

$$\beta = \sqrt{\frac{2cu_n\rho_1\lambda}{z\,l\,\rho_2}},$$

где u_n — температура поверхности, c и — λ теплоемкость и температуропроводность снега, l — влагосодержание, а ρ_1 и ρ_2 — соответственно плотность промерзшего и влажного участков снега.

X. ВЛАЖНОСТЬ СНЕГА, ТАЯНИЕ И ИСПАРЕНИЕ СНЕЖНОГО ПОКРОВА В ГОРАХ

В 1949-1950 г.г. было разработанно два метода определения влагосодержания снежного покрова: калориметрический и механический (24).

Калориметрическим методом влажность снега определяется с помощью измерения скрытой теплоты плавления, механическим же — с помощью замера изменения относительного об'ема воды при затвердевании.

При полевых исследованиях применялся калориметр адиабатического типа, роль адиабатической оболочки которого выполнял сосуд Дюара. Оптимальные условия работы калориметра следующие.

1. Интервал допустимого значения изменения температуры жидкости калориметра не должен превышать 20 С°.

2. Начальную температуру жидкости калориметра следуёт подбирать по возможности из условия: 2) $(\bar{u} - u) = \Delta u$,
где $\bar{u}$ — средняя температура окружающей среды.

3. Наилучшее соотношение между весом образца и количеством воды в калориметре находится из условия:

$$\frac{m}{\mathrm{M}+m} = \frac{\Delta u}{\mathrm{L}+u},$$

где $м$ — масса образца, M — масса жидкости калориметра. При этих условиях относительная погрешность измерения влагосодержания определяется:

$$\frac{\delta l}{l} \cong \frac{1}{l}\,\frac{\mathrm{L}+u}{\mathrm{L}}\,\frac{\delta\Delta u}{u}.$$

В литературе имеется указание (45) на возможность определения влагосодержания посредством электрического нагрева образца. Измерения в этом случае

должны сильно упроститься, т.к. температура образца в процессе таяния все время остается при 0° С., и легко создать адиабатические условия работы калориметра простым погружением сосуда в массу влажного снега. Сосуд для взятия проб может одновременно служить и калоприметром. Регистрации подлежат величины, входящие в выражение Джулева тепла ($Q = 0{,}249\ I^2Rt$), причем окончание таяния образца устанавливается без труда по моменту начала под'ема температуры в приборе. Отсчитывать же температуру нет необходимости. Высказывалось предположение (45), что данный метод может дать наиболее точные результаты. Однако детальный анализ убеждает нас, что при отсутствии тепловых потерь оба описанных метода примерно эквиваленты по своей точности.

Исследования над влажностью снега показали, что последняя не является однозначной функцией температуры снега, а зависит также от состояния снега до проведения опыта.

Если снег долгое время находится при 0° и пропитался водой, а затем охлаждается, то через 16 часов от начала опыта при понижении температуры до — 4° — — 5° в нем остаются заметные следы воды, повышающие его теплоемкость до 0,65.

Если же снег нагревается от — 8° — — 10° даже до температуры — 0,5°, то его влажность будет равна 0.

При температурах воздуха, варьирующих около 0°, поверхность снега сильно увлажняется, причем величина влажности может достичь 5-10 %.

Влажность снега резко изменяет его физические свойства как-то: теплопроводность, силу сцепления и т.д. Поэтому и эти физические свойства снега также являются многозначными функциями его состояния (ρ и u) при температурах близких к 0°.

К сожалению, в настоящее время дать какую-либо количественную зависимость между влажностью снега, его температурой в момент проведения опыта и ходом температуры до опыта не представляется возможным.

Необходимо отметить одно интересное явление, связанное с миграцией паров, обуславливающее как изменение структуры в толще покрова, а также, вероятно, и перенос влаги между склонами различной экспозиции. В 5 металлических кювет (диаметр — 7.2 см, высота — 6 см.) с сетчатым дном насыпался снег, затем кюветы взвешивались в термобарокамере при отрицательной температуре на аналитических весах и устанавливались одна на другой. Нижняя кювета имела металлическое дно, а верхняя — закрывалась металлической крышкой. Стойка из 5 кювет помещалась в войлочный цилинд для лучшей теплоизоляции. На крышку верхней кюветы устанавливалась электрическая печка, мощностью 20 W. Периодическим включением этой печки температура верхней кюветы поддерживалась — 2°, а нижняя кювета находилась при температуре — 12°.

По прешествии определенного времени все кюветы взвешивались вновь, и по изменению веса снега в кюветах делался вывод о миграции водяного пара.

Эти опыты подтверждают наличие миграции водяного пара в снегу при наличии температурного градиента, причем наибольшее изменение веса наблюдается у 5-й и 4-й квювет. Это можно об'яснить тем, что здесь вероятно, создается наибольший температурный градиент. Во всех случаях уменьшение веса снега больше, чем увеличение, т.е. испарение преобладает над конденсацией. Возможно это об'ясняется тем, что кюветы были недостаточно плотно пригнаны одна к другой.

По этим данным был расчитан коэффициент диффузии водяного пара снежного покрова, который в системе CGS оказался порядка $D = 10^{-1}$.

В результате исследования распределения температуры, изменения стратиграфии и послойной плотности в толще снежного покрова в полевых условиях было подтверждено, что температурный градиент обуславливает миграцию водяного пара из нижних, более теплых горизонтов снега в вышележащие слои. в некоторых случаях количество перенесенной мыссы составляло 0,33 г/сутки с одного см² при перепаде температур между верхним и нижним слоями порядка 15°-20° и глубине покрова 45 см, что неплохо созпадает с данными, полученными в лабораторных условиях.

Изменение массы снега на южном и северном склонах показали, что увеличение массы снега на 1 см² поверхности на северном склоне с 15 декабря по 20 января составляло 0,02 г/сут., на южном склоне за тот же промежуток времени масса снега уменьшалась, в основном, за счет испарения на величину

0,07 г/сут. Эти данные в какой то степени подтвердили высказанную раннее гипотезу о перекачке влаги со склона южной экспозиции на склон северной экспозиции (35). Процессы испарения с поверхности снега и конденсации на поверхности снега зависят от разности температуры поверхности покрова и приземного слоя атмосферы и меняются в зависимости от времени суток, года, экспозиции склона и характера залегания снега. В какой — то степени эти процессы отображаются уравнением Пузанова (27, 28).

XI. УРАВНЕНИЕ ТЕПЛОВОГО БАЛАНСА СНЕЖНОГО ПОКРОВА

Снежный покров представляет из себя прослойку, изолирующую почву от тепловых процессов, протекающих в атмосфере. При изучении теплового баланса снежного покрова необходимо учитывать, помимо теплообмена на границах раздела воздух-снег и снег — почва, тепловую энергию, израсходованную на фазовые превращения в покрове и верхних слоях почвы, а также поглощение лучистой энергии в толще снега. Турбулентный теплообмен атмосферы с поверхностью снега, как известно, выражается следующей формулой:

$$R_n - K_\tau C_p \rho_b \left(\frac{\partial u}{\partial z}\right)_b - K_\tau R \rho_b \left(\frac{\partial e}{\partial t}\right)_b - \int_0^{-h_1} \left\{[C_c \rho_c + R f_t(u)]\left(\frac{\partial u}{\partial t}\right)_c + L\left(\frac{\partial e_1}{\partial t}\right)_c + (1 - A) I_n l^{-K' \rho_c z}\right\} dz - \int_{-h_1}^{-h_2} \left[C_n \rho_n \left(\frac{\partial u}{\partial t}\right)_n + L\left(\frac{\partial e_1}{\partial t}\right)_n\right] dz + K_n \left(\frac{\partial u}{\partial z}\right)_n = 0$$

где K_τ — коэффициент турбулентного перемещивания, C_p — теплоемкость воздуха при постоянном давлении, ρ_b — его плотность, $\left(\frac{\partial u}{\partial z}\right)$ [градиент температуры в приземном слое воздуха.

Теплота, израсходованная на испарение или конденсацию на поверхности снега, равна

$$K_\tau R \rho_b \left(\frac{\partial e}{\partial z}\right)_b,$$

где R — скрытая теплота испарения, а $\left(\frac{\partial e}{\partial z}\right)_b$ — градиент влажности в приземном слое.

В продолжении двух зимних сезонов (1955-1956 г.г. и 1956-1957 г.г.) в Эльбрусской экспедиции ИПГ проводилось систематическое определение всех членов теплового баланса (36).

В результате проведенных исследований представляется целесообразным разбить радиационный баланс на два части в зависимости от длины волны падающей радиации I_n на инфракрасную часть R_u и видимую часть R_b.

Инфракрасная радиация поглащается первым 2-3 сантиметровым слоем снега, видимая же часть радиации проникает в толщу покрова до глубины 25-30 см. В результате поглощения видимой части солнечной радиации в снежном покрове в некоторых случаях наблюдалось повышение температуры на 3-4° С в средних слоях, о чем будет сказано ниже.

Этот член радиационного баланса можеть быть представлен в виде:

$$\int_0^{-h_1} A I_n l^{-K^1 \rho_c z} \lambda z, \qquad (16)$$

где A — альбедо снега, K^1 — коэффициент поглощения радиации, ρ_c — плотность снега и h_1 — его мощность.

Не трудно видеть, что интеграл (16) равен поглощенной радиации I_n-I_o, такая форма записи лучше отображает физический смысл процесса поглощения в снежном покрове.

Тепло, необходимое для изменения температуры в толще покрова можно представить в виде:

$$\int_o^{-h} C_c \rho_c \left(\frac{\partial u}{\partial t}\right)_c \lambda z$$

где $C_c\ \rho_c$ — об'емная теплоемкость снега, а $\left(\frac{\partial u}{\partial t}\right)_c$ — изменение темпаратуры в толще покрова. Теплота, выделяемая в результате замерзания или таяния снежных зерен, равна

$$L\int_o^{-h} \left(\frac{\partial e_1}{\partial t}\right)_c \lambda z$$

где $\left(\frac{\partial e_1}{\partial t}\right)_c$ — изменение жидкой фазы воды в покрове. Теплота, выделенная в результате сублимации о снежном покрове, может быть представлена в виде:

$$R\int_o^{-h^1} \left(\frac{\partial e^2}{\partial t}\right)_c \lambda z$$

где $\left(\frac{\partial e_2}{\partial t}\right)_c$ — изменение газообразной фазы воды в покрове.

Поскольку водяной пар в снежном покрове находится в состоянии насыщения, e_2 можно представить в виде $e_2 = f(u)$.

В первом приближении эту зависимость можно установить по формуле Магнуса (33).

Теплоту промерзания почвы, изменение температуры промерзшего слоя и поток тепла на границе промерзшего слоя можно учесть выражением вида:

$$\int_{-h_1}^{-h^2} \left[L\left(\frac{\partial e_1}{\partial t}\right)_n + C_n \rho_n \left(\frac{\partial u}{\partial t}\right)_n \lambda z + K_n \left(\frac{\partial u}{\partial z}\right) h_2 \right],$$

где $\left(\frac{\partial e_1}{\partial t}\right)_n$ — замерзшая часть воды в почве, $C_n \rho_n$ — об'емная теплоемкость почвы, $\left(\frac{\partial u}{\partial t}\right)_n$ — изменение температуры промерзшего слоя почвы, K_n — теплопроводность почвы, h_2 — граница промерзания.

На основании высказанного уравнения полного теплового баланса снежного покрова принимает следующий вид:

$$R_n - K_\tau C_v \rho_b \left(\frac{\partial u}{\partial z}\right)_b + K_\tau R \rho_b \left(\frac{\partial e}{\partial t}\right)_b +$$

$$+\int_o^{-h} \left\{ \left[C_c \rho_c + R f_t^1(u) \right] \left(\frac{\partial u}{\partial t}\right)_c + L\left(\frac{\partial e}{\partial t}\right)_c + A I_n l^{-K^1 \rho_c z} \right\} \lambda z +$$

$$+\int_{h_1}^{-h_2} \left[C_n \rho_n \left(\frac{\partial u}{\partial t}\right)_n + L\left(\frac{\partial e^1}{\partial t}\right)_n \right] \lambda z + K_n \left(\frac{\partial u}{\partial z}\right)_n = 0 \qquad (9)$$

Теплоту, выделяемую при конденсации или испарении влаги и почве можно не учитывать.

Во время таяния снега $\left(\frac{\partial u}{\partial t}\right) = 0$ и выражение (9) упрощается. В случае, когда снег долгое время находится при отрицательных температурах $\left(\frac{\partial e}{\partial t}\right)_0 = 0$

ОСОБЕННОСТИ ТЕМПЕРАТУРНОГО РЕЖИМА СНЕЖНОГО ПОКРОВА НА КАВКАЗЕ

Исследование температурного режима снежного покрова, а также радиационного баланса и теплообмена снега с воздухом и подстилающей поверхностью, (1953-1957 г.г.) позволили сделать некоторые выводы о характере теплообмена в толще снежного покрова и его теплового режима.

В середине зимы турбулентный теплообмен между приземным слоем воздуха и поверхностью снежного покрова в условиях Большого Кавказа исчезающе мал по сравнению с остальными компонентами теплового баланса. Из всех случаев расчета величина суточного теплообиена не превышает десятых долей малой калорий на см² поверхности снега.

Радиационный баланс составляет около 15-20 кал. на см² в сутки. Выделение скрытой теплоты фазовых переходов за счет миграции паров, возникающей вследствие температурного градиента между верхними и нижними горизонтами покрова, может в отдельных случаях за 4-8 часов привести к выделению тепла до 2х кал. в вертикальном столбике снега с основанием в 1 см² и высотой в 5-10 см. Для покрова в целом, за иключением поверхностного слоя, этот процесс является адиабатическим и приводит к уменьшению температурных градиентов.

Испарение и конденсация водяного пара на поверхности снега могут привести к выделению или поглощению около 10-20 кал/см². Поток тепла от почвы в толщу снежного покрова всю зиму сохраняет одно и то же направление снизу вверх и, как правило, не превышает 1-3 кал/см² в сутки. Таковы основные компоненты теплового баланса снежного покрова. Значение каждого из них меняется от погодных условий в больших пределах не только по величине, но и по знаку/например, радиационный баланс, конденсация и испарение с поверхности снега и т.д./.

Как увидим ниже, все вышесказанное необходимо учитывать при прогнозировании температуры в толще снежного покрова, ибо механизм теплообмена в толще покрова, а, следовательно, и температура последнего зависит от того, за счет какого компонента баланса снег приобрел или потерял тепло.

а) Теплообмен в толще снежного покрова в результате радиационного баланса снега.

Солнечная радиация, проникая в зону покрова, поглощается первым 30-и сантиметровым слоем снега, поэтому нагрев в результате солнечной радиации происходит по экспоненциальному закону одновременно до глубины 20-30 см. Непосредственные наблюдения над поглощением солнечной энергии в различных горизонтах снега (1956-1957 г.г.) показали, что на глубине 20-25 см. за счет этой энергии температура снега может повысится в полуденные часы на 2-3 градуса.

Если принять, согласно Вейнбергу Б. П. и др. исследователями, что снежный покров по своим радиационным свойствам близок к абсолютно черному телу, то следует ожидать, что охлажение снежного покрова за счет его излучения в ночные часы протекает также по экспоненте одновременно до глубины 25-30 см.

Нагрев снега и его охлаждение в результате поглощения и излучения радиации будет происходит почти одновременно до глубины 30 см. по закону экспоненты и, следовательно, на графиках термоизоплет будет отсутствовать запаздывание температурных амплитуд, наблюдаемых в почве. Этот вывод хорошо согласуется с ходом термоизоплет.

б) Теплообмен в толще снежного покрова, обусловленный фазовыми переходами на поверхности снега.

Совершенно иная картина температурного хода наблюдается в тех случаях, когда охлаждение или нагрев поверхности снега происходит в результате конденсационных процессов или испарения с поверхности. Так, в промежутке между 2/1 и 5,1 1957 г. создались погодные условия, характеризующиеся умеренным ветром (1-3 с/сек.) и таким соотношением температур поверхности снега и воздуха, при котором средняя удельная влажность воздуха 0,2-0,3 мб),

образовала относительную влажность у поверхности снега в 60-70 %. В результате этого с поверхности снега происходило интенсивное испарение и температура поверхности снега опускалась до — 28°.

Интересно отметить, что в этом случае колебания температуры в 2-3° проникали в толщу покрова до глубины не свыше 7-10 см. с запаздыванием на 7-9 часов. На глубине 30 см. за трое суток температура опустилась всего лишь на 1°-1,5°.

Аналогичная картина наблюдалась 24 декабря 1956 года и в другие дни. Таким образом, при изменении температуры снега, вызванном непосредственным выхолаживанием или нагревом его поверхности, колебания температуры проникают в толщу покрова на глубину 7-10 см. в сутки и экстремумы температур запаздывают на этой глубине на 7-9 часов.

Все выше сказанное необходимо принять во внимание при прогнозировании температур в толще снежного покрова, и поэтому метод Дюбюка и Монина (10) для прогнозирования температуры в толще снега можно применять в условиях Кавказа с большой осторожностью.

НЕКОТОРЫЕ ЗАМЕЧАНИЯ О ЛАВИННЫХ ПРОЦЕССАХ НА КАВКАЗЕ

Вопрос, связанный с лавинами, подробно освещен в литературе (5, 38, 42, 44, 48, 51, 52).

Поскольку до последнего времени и литературе не было согласованности в определении лавинного процесса, мы под лавиной будем понимать движение снежных масс по склону под действием силы тяжести, при котором: а) в движении участвуют большие массы снега, причем лавина возрастает за счет увлечения лежащих по пути снежных масс, б) скорость движения снежных масс превышает 7 м/сек, в) перемещение снежных масс происходит путем скольжения и качения снега по подстилающей поверхности, а также, в некоторых случаях, путем переноса по воздуху плотного облака снежной пыли, д) движение снежных масс сопровождается своеобразным свистом низкого тона (в случае сухого снега) или скрежетом (в случае мокрого снега), а иногда оглушительным шумом наподобие раскатов грома (в случае образования воздушной волны).

В настоящей работе лавины классифицируются по частоте схода и составу переносимого снега. 1 класс — систематические лавины, сходящие по одному и тому же руслу по несколько раз в год (но не реже одного раза в 3-5 лет). Этот класс лавин хорошо известен местному населению: участки, перекрываемые ими, легко различимы и поэтому, как правило, от таких лавин всегда можно уберечь сооружения. 2 класс — случайные или спорадические лавины, сходящие по одному и тому же логу не чаще одного раза в пять и больше лет. Эти лавины вызваны особыми, редкими для данного района метеорологическими условиями, имеющими место в большинстве случаев во время или сейчас же после снегопада (34).

Классы делятся, в зависимости от переносимого лавиной снега на 4 типа. I тип — лавины свежего, сухого, неуплотнившегося, пылевого снега. II тип — лавины смешанного снега, состоящего как из свежего, сухого, так и из уплотнившегося снега, в некоторых случаях слегка увлаженнного. III тип — лавины уплотнившегося сухого и слегка увлажненного снега. IV тим — лавины мокрого снега.

Класс случайных спорадических лавин состоит, главным образом, из лавин первого и второго типов.

Для лавин 3 и 4 типов Гоффом, Оттеном (6) и Саатчяном (8) (29), была выведена зависимость между углом наклона местности α, глубиной снежного покрова h, его удельным весом $g\rho$, коэффициентом трения f, длиной оторвавшегося пласта l, силой сцепления c и сопротивлением разрыву c_1, имеющая следующий вид:

$$h = \frac{\frac{c}{g\rho}}{\sin\alpha - f\cos\alpha - \frac{c_1}{g\rho e}} \qquad (10)$$

« Угол безопасности » — α_o — минимальный угол местности, при котором еще может сорваться лавина, независимо от глубины снега, определяется выражением:

$$\cos \alpha_o = \frac{\sqrt{1-f^2}-fc}{(1-f^2)g\rho l} \tag{11}$$

Вводя угол внутреннего трения φ и силу сцепления по бокам отрыва пласта c_2, предыдущие уравнения мы привели к следующему виду:

$$h = \frac{c_1 \tau l \cos \varphi}{g\rho l \tau \sin(\alpha - \varphi) - (l\tau + 2 c_2 c_\Delta \cos \alpha)}, \tag{12}$$

где τ — площадь срывающегося пласта, и

$$\alpha_o = \varphi + \arcsin\left[\frac{c_1 \tau + 2c_2 l}{g\rho l \tau} \cos \varphi\right] \tag{13}$$

Формулы (10, 11, 12, 13) определяют условия устойчивости уплотнившегося снега на горных склонах. При подстановке значений параметров, характеризующих снежный покров Кавказа, для α_0 получается минимальное значение в 14° (табл. 8).

Таблица 8

Тип снега	α_o
Сухой свежий снег	14,8°
Сухой Уплотнен. снег	23,5°
Мокрый снег	16,5°
Мокрый уплотн. снег	27,2°

Наблюдения над снежными лавинами на Кавказе показали, что в некоторых, исключительных случаях, при сухих снегопадах лавины сходили со склонов с уклоном в 10°-12°.

В результате обследования места срыва сухих лавин с некрутых склонов было установлено, что в этом случае снежные лавины образуются не путем отрыва пласта от подстилающей поверхности, а путем осыпания призмы рыхлого снега вдоль плоскости обрушения, проходящей в снежном покрове (7, 8).

Условия устойчивости сыпучего снега для уступа местности аналогичные условиям (26) — (27) имеют вид:

$$h_y = \frac{2c \sin(s-\alpha)\cos\varphi}{g\rho \sin^2 \alpha} \tag{14}$$

где α и β углы наклона склонов.

Для угла наклона плоскости обрушения ψ_{oy} имеем

$$\psi_{oy} = \frac{\beta - (2\alpha - \varphi)}{2} \tag{15}$$

для перегиба склона

$$h_n = \frac{2c \cos\dfrac{\beta-\alpha}{2} \cos\varphi}{g\rho \sin^2\left[\dfrac{90^\circ - \varphi}{2} + \dfrac{\alpha+\beta}{2}\right]} \tag{16}$$

и

$$\psi_{on} = \frac{180^\circ - (\beta + 2\varphi - 3\alpha)}{4} \tag{17}$$

Расчет скорости движения лавины без учета захвата лавиной лежащих на пути снежных масс приводит к занижению ее величины по сравнению с данными наблюдениями. Выражение для скорости лавин при этом имеет вид (8, 9, 31):

$$v_\alpha = (v_0 - v^*) l^{-kt} + v^* \tag{18}$$

где v^* — скорость равномерного движения лавины,

$$v^* = \frac{g}{K} \frac{\sin(\alpha - \varphi)}{\cos\varphi} \tag{19}$$

Здесь K — коэффициент лобового сопротивления движению (принималось, что лобовое сопротивление пропорционально первой степени скорости движения). В предположении, что сопротивление движению пропорционально квадрату скорости с учетом захвата лежащих на пути движения снежных масс (6, 7) для участка склона с углом наклона α_i скорость движения лавины $v_{\alpha i}$ равна.

$$v_{\alpha i} = \sqrt{\frac{\alpha_i}{b+1}\left[1 - \left(\frac{x^{(i-1)}}{x}\right)^{b+1}\right] + \left(\frac{x_o^{(i-1)}}{x}\right) v^2_{i-1}} \tag{20}$$

где $\alpha_{ix} = \frac{2g\sin(\alpha_i - \varphi)}{\cos\varphi}$, $x_o^{(i-1)} = \frac{K_1 h}{H} \sum_{n=1}^{i} x_n$,

$$b = 1 + \frac{K_1 H}{K_2 \rho h}$$

H — высота фронта лавины; h глубина снежного покрове, K — коэффициент сопротивления движению, K_2 — коэффициент захвата, v_{i-1} — скорость на предыдущем участке, x_n — протяженность участка с углом наклона α_n. С помощью формулы (20) можно вычислить скорость лавины на любом участке, если известна глубина покрова, характер отложившегося снега и уклон местности.

Расчет давленияе P при ударе лавины о неподвижное препятствие производится по формуле:

$$P = 10^{-2} \frac{\rho_1 \rho_2}{\rho_2 - \rho_1} v^2 \tag{21}$$

где ρ_1 и ρ_2-плотность снега до и после удара, $\Delta\rho = \rho_2 - \rho_1$, v — скорость движения лавины в момент удара.

Для получения численной величины P, необходимо рещать уравнение (21) совместно с уравнением, выражающим зависимость между ρ_1, ρ_2 и ρ, которая представляется формулой (2). Совместное решение уравнения (21) и (2) приводят к следующей формуле:

$$P = \frac{K_\varrho \rho_1 v^2 + \sqrt{(K_\varrho \rho, v^2 + 400 K_\varrho \rho_1 v^2}}{200 K_\varrho} \tag{22}$$

где $K_\varrho = 3{,}10 - 7{,}81\ \rho$.

Коэффициенты в уравнении (22) подобраны таким образом, что если ρ выражается в г/см^3, v — в м/сек, то давление в кг/см^2.

В 1954 г. было установлено 11 щитов, регистрирующих силу удара и 6 створов, записывающих скорость движения лавины.

Электрическая регистрация производится путем записи на осциллографе изменения силы тока, вызванного увеличением индуктивного сопротивления катушки в результате вдвигания в нее сердечника, соединенного с осью щита, воспринимающего удар лавины.

Из 12 наблюдений над самыми различными по характеру движения лавинами выяснилось, что расчетные формулы раза в 4 превышают наблюдаемую силу удара, последнее по всей вероятности связано с неточностью в определении коэффициентов K_1 и K_2 в уравнении. Отношение между рассчитанной и измеренной силой удара оказалось величиной постоянной (до 10 %), что указывает на правильность предпосылок, положенных в основу вывода формул.

На основании расчетов (7, 8) и экспериментальных данных, была рассчитана сила удара лавины в т/м² о неподвижное препятствия в зависимости от глубины оторвавшегося снежного пласта (h), высоты фронта лавины (H), угла наклона склона (α) и длины пройденного пути (S). Эти данные приведены в таблице 9.

Таблица 9

h = 1 метр H = 3 метра

S в метрах

α_o	500	750	1000	1250	1500
20	2,26	2,37	3,3	3,8	4,2
25	6,5	8,45	10,2	11,9	13,3
30	9,6	12,7	15,7	18,5	21,3
30	9,6	12,7	15,7	18,5	21,3
35	12,4	16,6	20,7	24,6	21,5
40	15,0	20,3	25,4	30,4	35,3

h = 2 метра H = 3,5 метра

S в метрах

α_o	500	750	1000	1250	1500
20	2,5	3,1	3,7	4,2	4,7
25	7,2	9,4	11,5	13,4	15,3
30	10,8	14,3	17,4	21,0	24,3
35.	13,6	18,5	23,5	28,0	32,5
40	16,9	23,0	25,5	34,7	40,5

h = 3 метра H = 5 метров

S в метрах

α_o	500	750	1000	1250	1500
20	2,5	3,1	3,7	4,2	4,7
25	7,2	9,4	11,6	13,6	15,5
30	10,9	14,5	18,0	21,3	24,6
35	14,0	18,6	23,5	28,5	38,0
40	17,2	23,6	29,5	35,4	41,3

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LE CHANGEMENT DE LA DENSITÉ DE LA NEIGE EN HONGRIE

M. KÉRI* & P. SALAMIN**

Résumé

La connaissance de la densité de la neige est indispensable non seulement pour la détermination de la teneur en eau de la neige recouvrant le sol, mais aussi pour la solution de nombreux autres problèmes hydrologiques.

La présente étude se propose d'examiner le changement de la densité de la neige dans l'espace et dans le temps et d'analyser la nature de ce changement, sur la base d'observations faites sur place en plusieurs endroits des régions montagneuses et des plaines de la Hongrie, pendant les années 1954-1956.

L'élaboration des données des observations a permis d'étudier le caractère du changement de la densité et de reconnaître un certain nombre d'influences. Aussi, au cours des recherches poursuivies par rapport à l'espace, l'on a pu constater que la valeur de la densité ne change que dans une mesure relativement faible, même sur un terrain de relief très varié (en valeur absolue elle ne dépasse pas 0,1 kg m^{-3}). L'influence du relief (surface, point cardinal, altitude) et celle de la végétation peuvent être démontrées dans beaucoup de cas, tandis que dans d'autres, elles paraissent insignifiantes à côté des influences climatiques générales ou ne sont même pas perceptibles. Au cours des recherches faites par rapport au temps, on est arrivé à établir un procédé simple qui tient compte du rapport existant entre la densité et l'épaisseur de la neige aussi bien que de la connexion de leur changement, il permet d'étudier le problème du changement par rapport au temps dans les conditions climatiques de la Hongrie. En même temps l'on a déterminé les valeurs extrêmes de la densité pour la Hongrie. Les influences climatiques ont été prises en considération avec soin dans chacun des cas.

Pour déterminer l'équivalent en eau de l'enneigement et pour résoudre un grand nombre de problèmes hydrologiques, il est indispensable de connaître les changements de la densité (*) de la neige.

Le changement de la densité de la neige, dans son essence, est le résultat des processus d'économie d'énergie thermique et des processus d'économie d'eau qui s'opèrent dans l'enneigement et forment la structure cristalline granulée ou amorphe. Parmi les *facteurs d'économie d'énergie thermique*, notons les plus importants :

la condensation ou la transformation en état liquide de la teneur en vapeur d'eau contenue dans les interstices et sur la surface de la couche de neige,

la radiation solaire,

la radiation de l'air dans la direction de la couche de neige,

l'échange de chaleur entre la couche de neige et l'air,

l'influence de la pluie tombée sur la couche de neige,

la conduction de la chaleur à partir de l'air et du sol.

Dans le domaine de l'économie d'eau, l'on peut considérer comme des facteurs essentiels :

le rôle des précipitations en forme de pluie et de neige.

(*) Institut Météorologique, Budapest.

(**) Chaire d'Hydrotechnique N° 1, Université des Sciences Techniques, Budapest.

(*) Pour noter l'neige et faciliter l'exposé, nous employons le terme « densité » pour désigner le poids volumétrique apparent de la neige à l'état de structure naturel, c'est-à-dire le poids d'un volume de neige égal à l'unité que l'on mesure directement sur l'échantillon prélevé.

Fig. 1 — Lieux et stations de mesure.

le rôle de la teneur en vapeur d'eau qui se transforme en état liquide ou solide dans la partie intérieure ou à la surface de la couche de neige,

l'évaporation,

le ruisselement, l'écoulement et l'infiltration de la neige fondue.

Conformément à ce fait, et pour étudier le problème d'une façon précise, il nous faut examiner séparément chaque processus de l'économie d'énergie thermique et de l'économie d'eau, et les considérer aussi dans leur ensemble, telle que cela a été préconisée par plusieurs auteurs dans la littérature. Cependant dans la pratique, le plus souvent, il nous est impossible de soumettre ces facteurs — surtout ceux de l'économie d'énergie thermique — à une telle étude précise, puisqu'il nous manquent, par rapport à une étendue plus grande, les données pouvant servir de point de départ pour nos recherches. Ainsi, nous sommes obligés de recourir à une méthode à la fois plus simple et plus directe — et en même temps moins précise —, afin de déterminer le changement de la densité et d'éclaircir, avant tout, *la nature des changements que l'on peut observer dans l'espace et dans le temps.*

Ce fut cette voie que nous avons suivie *en Hongrie* pour étudier, au cours de ces dernières années, le changement de la densité de l'enneigement. Cette étude nous a conduit à plusieurs résultats par rapport aux changements intervenus dans l'espace et dans le temps. Nous en donnerons, dans ce qui suit, les plus importants. Nous décrirons d'abord les lieux où les recherches ont été faites et les méthodes que nous avons employées, et nous exposerons après les conditions générales dans lesquelles l'enneigement se forme en Hongrie

Le lieu et le temps de l'étude; les méthodes de recherches.

Nous avons étudié le changement de la densité de l'enneigement parallèlement dans les régions montagneuses, dans les plaines et dans la ville.

Les mesures les plus importantes effectuées sur la neige, *en régions montagneuses*, sont celles que nous avons faites dans les monts Bükk et dans les monts Mâtra.

Dans *les monts Bükk* (voir sur la Figure 1, numéro 7), les mesures d'enneigement ont été faites en 1954 dans le bassin versant supérieur du ruisseau du Garadna, qui s'étend sur 16,7 km² et s'enclave entre le plateau central du Bükk et celui dans la partie nord et nord-est de la montagne. La vallée érosive du Garadna, préformée par de jeunes ruptures, constitue la limite entre les deux plateaux. Le plateau central est, en grande partie, de l'étage moyen de la période triasique; il est fortement karstique et sa structure est constituée de calcaires gris. Une grande partie de son eau provenant des pluies et de la fonte des neiges, afflue, sous la terre, à travers un réseau d'eau karstique, vers la vallée du Garadna. La partie calcaire du plateau nord et nord-est, est de même, fortement karstique, tandis que les parties voisines du bassin versant du Garadna, formées principalement de schistes, sont coupées par des vallées d'érosion. Une partie de l'eau de ce plateau s'écoule également vers la vallée de la Garadna. *La surface du bassin exploré est découpée, ses pentes sont abruptes* et dans son coin sud-ouest se trouvent des dolines. *Sa vallée principale s'allonge dans la direction de ouest-est*, ce qui a permis de faire une étude précise sur les influences du relief aux versants exposés au nord et au sud. Son sol est de « redzina » et de terres forestières décolorées. Son revêtement naturel est de forêts de hêtres. Une faible partie (4,5 %) de l'étendue des plateaux élevés se trouve recouverte d'herbages. Sont point le plus bas est à 350 m au-dessus de niveau de la mer Adriatique, son point le plus haut atteint les 940 m d'altitude. La majeure partie de la neige fondue provenant directement du bassin versant étudié se rassemble sur la surface avant de se verser dans le Garadna, et ce n'est qu'une faible partie de l'eau qui s'infiltre dans la roche pour parvenir au ruisseau. [11]

Dans *les monts Matra* (4) les mesures ont été effectuées en 1955 et en 1956 sur le bassin versant de 44,8 km² du ruisseau « Kövicses ». Le bassin est constitué surtout par le plateau de Mátra dont l'altitude moyenne varie entre 700 et 800 m, tandis qu'une faible partie du bassin s'étend sur la limite du Mátra de l'ouest et du Nord. La partie formée par le plateau (sans compter quelques monts côniques qui en forment la bordure), *est une pénéplaine faiblement ondulée et aplatie.* Les vallées principales, s'éloignant de la région des sources vers l'Ouest, s'enfoncent profondément dans le sol et le découpe de plus en plus fortement. La partie basse du bassin versant, située à la limite des Mátra de l'Ouest et du Nord, sert de *transition entre une région de collines et le plateau.* A cet endroit se trouve une particularité caractéristique : *le secteur asymétrique du Kövicses.* Le point le plus bas du bassin versant est à 175 m, le plus haut à 946 m au-dessus de la mer. Sa structure fondamentale est faite de produits volcaniques, surtout de piroxène-andésite-tuf. Au pied ouest de la montagne, l'andésite cède la place à un système de cônes de déjection établi sur un fond de sédiment caillouteux et sablonneux. Le sol du bassin est un sol forestier, quelquefois brun ou décoloré. En bas, il est revêtu de forêts de chênes; aux plateaux les plus élevés et sur le versant exposé au nord, de forêts de hêtres. En beaucoup d'endroits, on trouve des prairies et des pâturages. Des champs labourable et des vergers n'apparaissent que dans les parties les plus basses, à l'Ouest. L'eau des pluies et de la fonte des neiges s'écoule sur la surface et descend vers la vallée du Kövicses.

Les mesures de neige faites dans les monts Mátra, ont été complétées par des observations recueillies au nord des Mátra, en 1955 et en 1956, dans le bassin versant du *ruisseau du Tarjan* (5), d'une étendue de 89,2 km² et en 1955 dans celui *du ruisseau du Bárna*[2] (6) dont la superficie couvre 42,0 km². Pour le premier, le point le plus bas est à 240 m et le point le plus haut à 625 m au-dessus du niveau de la mer; pour le second ils sont respectivement à 220 m et à 628 m.

Nous avons faits des recherches analogues aussi sur un autre terrain de caractère également montagneux : dans les *montagnes de Buda* (3) s'élevant à 280-490 m d'altitude.

L'étude de la plaine a été entreprise à quatre endroits différents. De 1954 à 1956, sur le terrain s'étendant *entre le Danube et la Tisza,* à la station « Komlósi »,[2] sur une clairière sablonneuse dans une forêt protégée des grands vents. *Au-delà de la Tisza,* à Szarvas [3] (10) et dans le bassin versant de la Vallée de Kondoros [3](11), champ d'expérience aménagé à l'étude d'économie d'eau, couvrant une étendue de 21,5 km² et ayant une altitude de 87,5 m au-dessus de la mer Adriatique. A cet endroit, sur des terres labourées et des sols d'argile des prairie, nous avons mesuré, en 1954, respectivement en 1956, la valeur de la densité de la neige et son changement. De semblables recherches ont été effectuées aussi à Taskony (9), également au-delà de la Tisza, en 1954, dans un verger, sur un sol de terre limoneuse.

Parallèlement à ces investigations poursuivies en régions montagneuses et en pays de plaine, nous avons recueilli des observations aussi à *Budapest,* notamment au jardin de l'Université des Sciences Techniques (1) et au parc de l'Institut Météorologique (2) entourés de bâtiments et de pavillons.

Au cours des recherches, on a déterminé d'abord *la densité de la neige* .On a pesé les échantillons de neige pour obtenir la densité ou plus exactement le poids spécifique apparent. *L'épaisseur de l'enneigement* a été mesurée en même temps.

Pour *prélever les échantillons* nous avons procédé soit en tenant compte du bassin versant tout entier et prélevant des échantillons tous les sept, éventuellement, tous

[2] Observations de l'Institut de Recherches de Ressources Hydrauliques, Budapest.

[3] Observations de l'Institut de Recherches d'Irrigation et de Culture de Riz, Szarvas.

les quinze jours, soit en déterminant la densité de la neige tous les jours, sur un certain nombre de points sélectionnés.

Dans *les recherches relatives aux bassins versants*, les endroits des prélèvements avaient été chosis de telle sorte que l'ensemble des prélèvements soit susceptible de donner les caractéristiques complètes du bassin et contienne toutes les influences possibles exercées par le relief et la culture, sans laisser prévaloir les unes au détriment des autres. On a donc fait des prélèvements d'échantillons également ([11]) :

dans des terrains différents du point de vue du relief et de la topographie (fond de vallée, versant, plateau, crête de montagne, vallées serrées et étendues, etc);

dans des terrains de situation différente par rapport aux points cardinaux (vallées et pentes exposées au Nord, au Nord-Ouest, à l'Ouest, etc.);

dans des terrains ayant une culture identique (prairie, pâturage, forêt, terrain déboisé et couvert de taillis, etc).

On s'est efforcé de réaliser également la simultanéité des prélèvements d'échantillons en les recueillant dans un intervalle de temps bref, avant même que des changements importants puissent intervenir dans la structure de l'enneigement. Les prélèvements réunie de la manière indiquée ont été contrôlés aux divers occasions de prélèvement, par l'approximation des résultats obtenus à l'aide des fonctions de répartition de la formule Pearson III. Sur la Figure 2, on a dessiné les courbes repré-

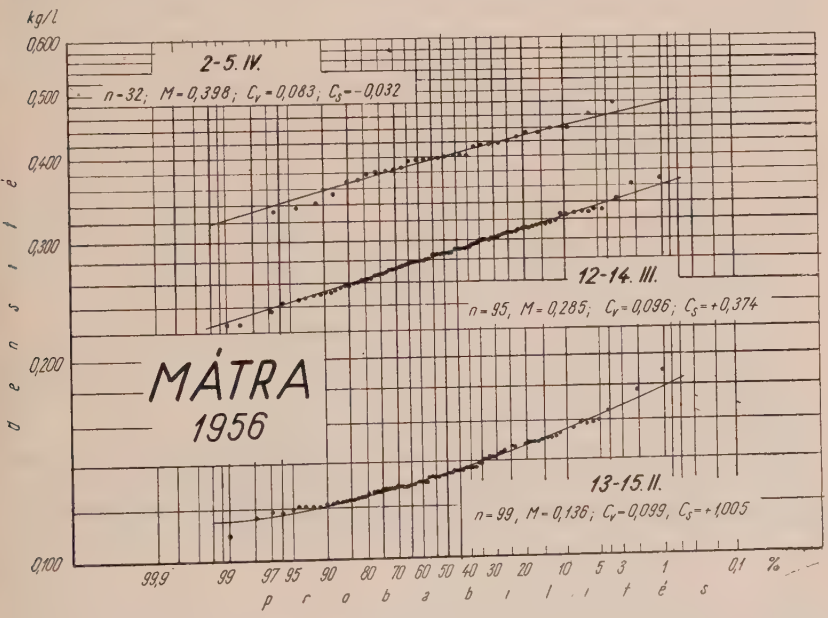

Fig. 2 — Contrôle des prélèvements d'échantillons à l'aide de la fonction de répartition.

sentant les fonctions de répartition de la formule Pearson III, et présenté aussi les points caractérisant les probabilités empiriques de chaque résultat d'observation pour trois occasions de mesures effectuées dans les Mátra en 1956 (13-15 II., 12-14 III. et 2-5 IV.). Comme on peut l'observer, les fonctions de probabilité théorique approchent très bien des valeurs des probabilités empiriques, ce qui prouve que les prélèvements d'échantillons avaient été bien disposés dans l'espace et dans le temps.

Nous avons fait des études sur toute l'étendue du bassin versant, dans les monts Bükk (7) et les monts Mátra (4), dans le pays de collines et de montagnes au nord des Mátra (5 et 6), dans la montagne de Buda (3), ainsi que dans le région d'au-delà de la Tisza, dans le bassin versant de la vallée Kondoros. Dans les *monts Bükk*, nous avons étudié le changement de la densité de la neige pendant quatre semaines, à 70 endroits différent s'en prélevant des échantillons tous les quinze jours. Nous avons cherché à déterminer surtout les influences exercées par le relief, le mode de la culture des terres et la végétation. Dans les *monts Mátra*, nous avons étudié le changement de la densité à 130 emplacements; en 1955, pendant quatre semaines, en 1956, pendant huit semaines, en prélevant des échantillons tous les huit jours. A l'aide de ces échantillons nous avons pu caractériser non seulement les changements qui s'opèrent sur la surface d'une pénéplaine faiblement ondulée, mais il nous a été possible de démontrer, en bien des cas, des influences morphologiques particulières aussi (influence des vallées encaissées, effet de l'asymétrie des vallées, etc.). *Dans les autres régions* mentionnées plus haut, nous n'avons effectué d'études couvrant toute l'étendue qu'exceptionnellement, pour des raisons de comparaison.

Au cours des examens quotidiennement faits *aux points d'observation ayant des caractéristiques climatiques bien définies*, on a fait plusieurs prélèvements par jour. Nous avons procédé de cette façon à Budapest, dans le jardin de l'Université des Sciences Techniques (1); dans le parc de l'Institut Météorologique (Office Météorologique) (2), dans la montagne de Buda, au parc de l'Observatoire astronomique (3), dans les monts des Mátra, à Mátraszentlászló (4), à la station « Komlósi » (8) entre le Danube et la Tisza et, enfin à Taskony (9) et à Szarvas (10), au-delà de la la Tisza. Les prélèvements recueillis sur ces points, tous les jours, nous ont servi *surtout* à caractériser les changements dans le temps.

Le nombre des échantillons de neige examinés pendant la durée de trois ans, se répartit comme suit :

	1954	1955	1956
en région montagneuse	904	1223	2372
en pays de plaine	937	66	396
en ville	548	210	197

Le nombre total s'élève, en rond, à 6850. L'étude d'un nombre d'échantillons aussi considérable nous a permis d'obtenir, par la déduction, un certain nombre de résultats touchant à la densité de la neige.

Parallèlement aux mesures de la densité et de l'épaisseur de la couche de neige, nous avons pris en considération les relevées d'observations *des stations météorologiques* relatives aux points et aux régions étudiées. A la station météorologique de Mátraszentlászló (1), nous avons mesuré continuellement, de jour en jour, la température au-dessus et en dessous de niveau de la neige.

Les conditions générales de la formation de l'enneigement (2-6)

On peut prévoir la formation de la couche de neige, en Hongrie, pour toute la période qui s'étend entre le premier et le dernier jour de la chute de neige, mais sa durée n'est pas, naturellement, aussi longue. *Le premier jour de la chute de neige* apparaît dans la deuxième moitié d'octobre au plus tôt, dans la région montagneuse dont l'altitude dépasse les 500 m au-dessus du niveau de la mer Adriatique. Au début de novembre, il y a de la neige dans toutes les montagnes d'altitude moyenne, et dans les premiers jours de la troisième décade de ce mois, la neige commence à tomber

presque partout en Transdanubie, et même dans la partie nord-est de la plaine hongroise, de l'Alföld. *La dernière chute de neige* — dont la date moyenne ne peut être calculée avec la certitude voulue à cause de l'incertitude psychologique qui entache les observations —, se situe entre la troisième décade du mois de mars et les derniers jours de la première décade de mai; elle arrive dans l'ordre inverse des régions énumérées.

Le premier jour calculé en moyenne où *la couche de neige se maintient déjà* est autour du 10 novembre (Mátra-Bükk). Cependant, vers le milieu de ce mois, la première couche de neige recouvre déjà toute la région montagneuse de la Hongrie. Dans la première moitié de décembre, la Transdanubie et la partie nord de l'Alföld se couvre aussi de neige, et dans la deuxième moitié du mois (jusqu'à 24 décembre au plus tard), la partie sud de l'Alföld est également enneigée. Comme *la disparition de la couche de neige* dépend, en Hongrie, beaucoup plus de la quantité de neige accumulée pendant l'hiver que de la dernière chute de neige (qui ne donne généralement qu'une couche de neige incapable de se maintenir), il n'est pas impossible que la date des deux phénomènes, calculée en moyenne, soit tout près l'une de l'autre, que l'enneigement se conserve même au-delà du dernier jour de neige calculé en moyenne. Dans tous les cas, dans le sud du pays qui est la partie la plus chaude, la couche de neige disparaît vers le milieu de février. Au mois de mars, il est rare de trouver de la neige dans les plaines hongroises. Mais la partie sud-ouest de la Transdanubie et les régions de montagnes moyennes restent recouvertes de neige jusqu'au milieu de mars; dans les hautes montagnes du pays, le manteau de la neige ne disparait que dans la deuxième moitié de ce mois, et recouvre les cimes les plus hautes jusqu'aux premiers jours d'avril.

Dans l'intervalle compris entre ces dates de limite, il y a des interruptions plus ou moins longues, et *le nombre des jours pendant lesquels l'enneigement continue à se maintenir*, n'est que les deux ou trois-cinquièmes de la totalité des jours. La couche de neige recouvre le sol pendant 30-45 jours à l'Alföld et de 35 à 50 jours dans la Transdanubie; tandis que dans les autres régions non-montagneuses, en fonction de l'altitude d'au-dessus de la mer, de 30 à 60 jours. Dans les montagnes de la Hongrie, la durée de la couche de neige est de 50 à 100 jours, à Kékestető (1015 m), la plus haute cime de la Hongrie, elle est de 104 jours.

En octobre, *la durée moyenne de la couche de neige* est de 1 à 2 jours, mais dans les régions montagneuses seulement, ailleurs la neige ne se maintient pas encore à cette époque. En novembre, la moyenne est de 0 à 5 jours à l'Alföld, de 5 à 9 jours dans les montagnes. Au mois de décembre, elle est de 5 à 15 jours pour la grande plaine et de 15 à 20 jours pour les montagnes. En janvier, elle est de 15 à 20 jours, respectivement de 20 à 30 jours. Au mois de février, on peut compter de 8 à 15 jours pour l'Alföld et de 15 à 18 jours pour les montagnes. En mars, la couche de neige ne se maintient plus que pendant 2 à 5 jours dans la grande plaine et de 5 à 15 jours dans les montagnes.Au mois d'avril, on ne trouve plus de trace de la couche de neige à l'Alföld, mais elle se maintient de 1 à 2 jours dans les régions périfériques du Sud-Ouest et du Nord, et de 2 à 5 jours dans les montagnes. En mai, la couche de neige est une rareté même dans les plus hautes montagnes du pays.

L'épaisseur moyenne de la neige, calculée pour toute la durée de l'hiver, est de 4 à 6 cm à l'Alföld, de 5 à 10 cm dans la Transdanubie, de 10 à 15 cm dans les montagnes plus hautes et de 4 à 8 cm dans les autres parties de la Hongrie. Bien que les jours d'enneigement soient les plus nombreux en janvier par rapport à la longueur du mois, les épaisseurs de neige maxima, calculées en moyenne, n'apparaissent qu'en février (à l'Alföld elles sont de 4 à 8 cm, dans la Transdanubie de 8 à 10 cm, dans la région montagneuse, aux niveaux moins hauts de 4 à 10 cm et dans les endroits plus hauts de 10 à 25 cm). Ce fait est significatif : elle montre que l'accumulation de la précipitation sous forme de neige continue à augmenter même en février et la couche de neige ne se fond, avec une grande rapidité, que vers

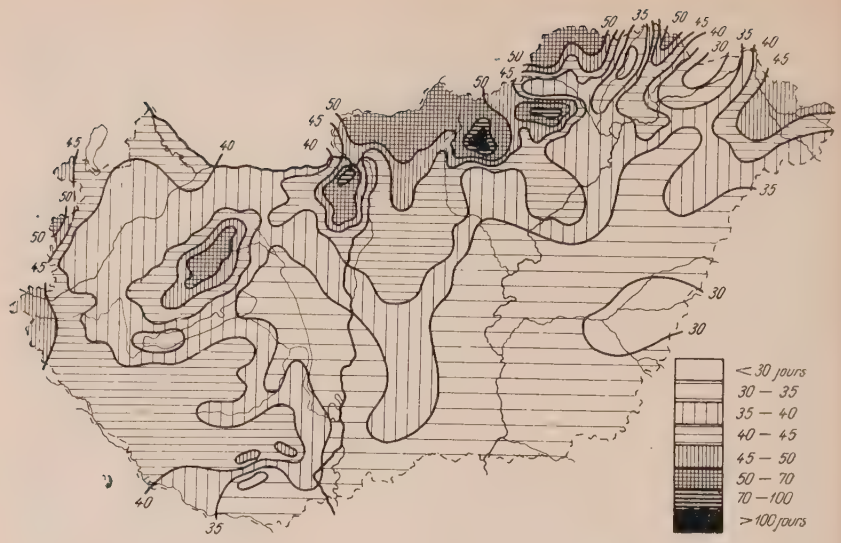

Fig. 3 — La répartition moyenne des jours d'enneigement en Hongrie.

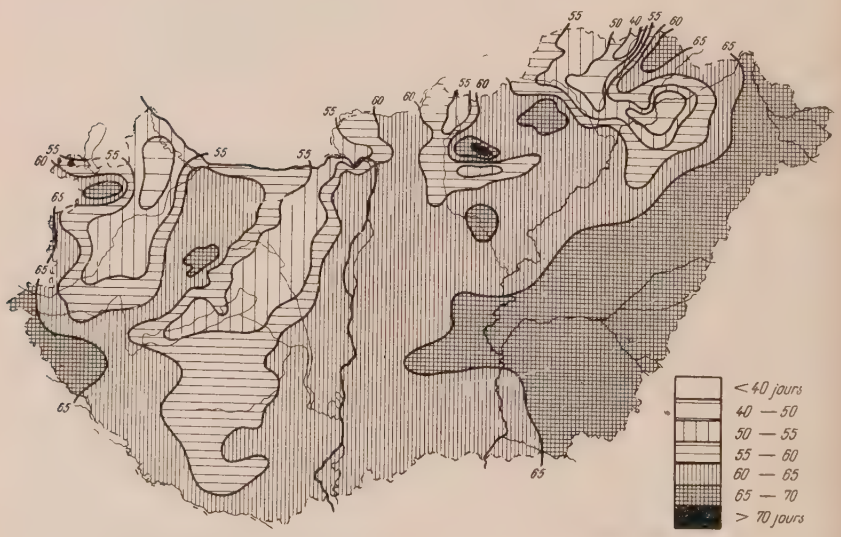

Fig. 4 — La répartition des jours d'enneigement en Hongrie pendant l'hiver 1953/54.

la fin de ce mois quand le temps commence à s'adoucir. Par suite de ce dernier fait, février est plus chaud, du point de vue de la température moyenne que le mois de janvier. Cependant, la couche de neige se maintient dans la majorité des cas, jusqu'à la fin de février :, et si l'on tient compte de tous les phénomènes climatiques qui caractérisent l'hiver, février peut être considéré, en Hongrie, comme le mois le plus hivernal.

C'est un trait caractéristique des conditions de neige en Hongrie, que *l'épaisseur maximum de l'enneigement ne dépasse jamais* 1 *mètre* sauf les hautes montagnes. Dans les montagnes où la précipitation totale est essentiellement plus importante que partout ailleurs dans le pays, il arrive assez souvent que l'épaisseur maximum de l'enneigement atteigne ou dépasse cette limite. Cependant, en général, on peut constater que l'épaisseur maximum de la couche de neige à l'Alföld reste au-dessous de 50 cm dans la périfërie d'est, et au-dessous de 50 à 75 cm dans les autres parties de la plaine, tandis que dans la Transdanubie et dans les montagnes, elle varie entre 75 et 100 cm.

Pour montrer *l'invariabilité relative des conditions de neige* dans le temps, en Hongrie, on a résumé dans les Tableaux I-II et sur les Figures 3-4 les résultats des observations effectuées dans les monts Mátra (à Kékestetö) au cours des 28 hivers derniers (1929/30-1956/57).

TABLEAU I

Fréquence des jours d'enneigement à Kékestetö par intervalles de 5 jours

mois	nombre des jours pendant lesquels l'enneigement se maintient					
	0-5	6-10	11-15	16-20	21-25	26-31
	fréquence en %					
X	96	—	4	—	—	—
XI	50	12	19	15	4	—
XII	11	12	19	15	8	35
I	—	—	4	4	8	84
II	—	4	4	4	12	76
III	19	19	12	19	4	27
IV	69	19	4	8	—	—
V	100	—	—	—	—	—

On peut constater après le Tableau I que l'enneigement qu'on a vu revenir le plus souvent pendant la période allant de décembre à mars et, dans plus que les trois quarts des cas examinés, en janvier et en février, c'était la couche de neige de durée de 26 à 31 jours. Comme il est indiqué dans le Tableau II, dans presque les deux

TABLEAU II

Fréquence de la durée d'enneigement à Kékestetö

durée de l'enneigement en jours					
0-25	26-50	51-75	76-100	101-125	126-152
fréquence en %					
—	—	12	27	42	19

tiers des cas examinés, l'on peut compter, dans les monts Mátra, avec un enneigement continu d'une durée qui atteint ou dépasse la moyenne. Les Figures 3 et 4 font bien voir que les traits principaux caractérisant la répartition des jours d'enneigement, sont les mêmes en Hongrie pour la moyenne des années et pour chaque année comptée séparément. La partie nord-est de la Hongrie (les vallées du Sajó, du Hernád et du Bodrog) où les precipitations de l'hiver sont les plus faibles, restent en arrière du point de vue du nombre des jours d'enneigement aussi par rapport aux autres régions du pays. Les territoires restant couverts de neige le plus longtemps, — par rapport à la moyenne aussi bien qu'en ce qui concerne les cas particuliers, — se trouvent dans la région montagneuse. Les valeurs caractéristiques des territoires dont les conditions de neige, pour certaines années, s'écartent sensiblement de la moyenne — comme cela fut le cas pour le bord Est de l'Alföld pendant l'hiver 1953/54, — restent en dessous des valeurs de la région montagneuse où les conditions de neige sont toujours les plus favorable.

Aux différences qui se manifestent entre les valeurs moyennes à l'intérieur des régions naturelles plus vastes (sur nos plaines par exemple, la différence du nombre des jours d'enneigement peut varier de 15 à 20 jours), correspondent, chaque année, à côté des valeurs moyennes plus basses ou plus hautes, selon que l'hiver avait été pauvre ou riche en neige, des différences de même ordre de grandeur. Il en est de même pour nos régions montagneuses, mais là, en conformité avec les différences d'altitude plus importantes, les différences de valeur moyenne sont plus élevées et les variations de la durée d'enneigement plus grandes. Comme on le voit, en Hongrie, *les variations de la durée d'enneigement dans le temps et dans l'espace montrent une identité rassurante.*

Si l'on examine les conditions générales de la formation de l'enneigement, on peut constater que le nombre des jours d'enneigement par mois, ainsi que sa durée moyenne et la moyenne et le maximum de son épaisseur *représentent*, conformément à la situation de géographie physique et climatique du pays, des valeurs relativement peu élevées et que la variation de ces valeurs dans l'espace et dans le temps reste relativement faible. *La valeur relativement réduite des caractéristiques de l'enneigement et leur variation* motivent suffisamment les résultats principaux qu'on a obtenus au cours des recherches sur la densité de la neige et qui vont être présentés dans les chapitres qui suivent.

Le changement de la densité dans l'espace

Pour déterminer *le changement général dans l'espace*, nous nous sommes servis des relevés d'observations recueillies dos les montagnes Bükk et Mátra et sur le territoire s'étendant au Nord de la montagne Mátra. Ces données confirment en tous points, le fait *que la valeur de la densité*, dans les conditions climatiques de la Hongrie, *ne change que dans une mesure relativement faible.* Les données démonstratives de cette constatation assemblées par la voie du calcul des probabilités, sont résumées dans le Tableau III et dans les Figures 5, 6 et 7. Dans ces Figures, afin de permettre d'embrasser d'un regard les recherches, nous avons appliqué un système de coordonnées à trois axes. Sur l'axe horizontal, on a représenté la densité. Sur l'axe vertical on a porté, en partie, les moments du prélèvement des échantillons qui devaient déterminer les axes horizontales des figures de détail, et en partie la fréquence juste par rapport à cet axe horizontale mentionné dernièrement. Les Figures ainsi présentées nous ont permis de mesurer non seulement les changements dans l'espace, mais dans le temps aussi. Si l'on confronte les Figures avec les colonnes correspondantes du Tableau III, on se rend compte du fait que *la différence des valeurs de densité de probabilité de* 10 % *et de* 90 % *non seulement ne dépassent pas*

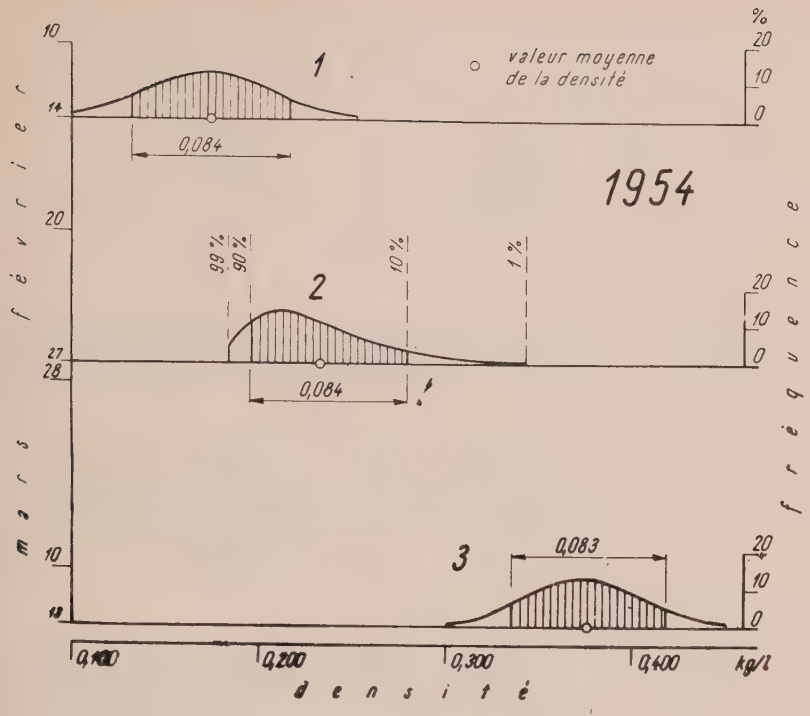

Fig. 5 — Étude de la répartition des valeurs de la densité dans l'espace et dans le temps. Mts. Bükk, 1954.

$\Delta\gamma = 0{,}12$, *mais restent, généralement bien en dessous de celui-ci, c'est-à-dire les densités des échantillons, pour* 80 % *des cas examinés, ne se différencient pas les unes des autres dans une mesure qui dépasserait* $\Delta\gamma = 0{,}12$ *dans aucune année et dans aucun zone étudié.* Ce résultat qui découle naturellement des causes mentionnées dans le chapitre précédent, signifie, en réalité, qu'en Hongrie la densité de la neige ne change que dans une mesure relativement faible.

Au cours de l'étude *des valeurs extrêmes*, on a constaté, que dans les conditions de Hongrie, la densité de l'enneigement frais est : $\gamma_{min} = 0{,}09\text{-}0{,}12$ kg/l. Nous avons observé, dans des cas particuliers, des couches de neige d'une structure considérablement moins compacte. Si la couche de neige est composé de neige mouillée fraichement tombée, sa densité est beaucoup plus grande. L'autre valeur extrême de la densité apparaît dans le processus final du dégèle. Dans le cas d'une couche de neige très mince dont l'épaisseur n'atteint que 10 ou 20 cm, la neige se fond déjà au-dessus d'une densité de $\gamma = 0{,}20$, et commence à suinter; dans le cas où l'enneigement est plus épais et, dans certains cas même s'il est moins épais, la fonte se termine lorsque la neige atteint la densité de $\gamma_{max} = 0{,}36\text{-}0{,}40$. La densité de la neige ne peut s'élever au-dessus de cette valeur extrême que difficilement, dans des conditions particulières et sur certains points du terrain, seulement dans les restes de couches de neige. Ces restes de couches de neige d'une densité élevée se rencontrent surtout à la fin de l'hiver, le plus souvent au long des murailles de loess exposées au Nord, dans les

Application du calcul des probabilités à l'élaboration des données recueillies

Année	N°	region étudiée	date des prélèvements d'échantillons	*n*	Cv	Cs	moyenne	1	3	5	
1954	1	Bükk (7)	II.12-15	88	0,189	+0.071	0,174	0,252	0,237	0,229	0,2
	2		II.26-28	101	0,150	+1.252	0,232	0,343	0,313	0,299	0,2
	3		II.12-14	30	0,086	+0,179	0,376	0,455	0,439	0,431	0,4
1955	1	Mátra (4)	II.29-III.1	47	0,102	+0,622	0,188	0,241	0,229	0,223	0,2
	2		III.7-8	95	0,180	+1,584	0,241	0,388	0,346	0,326	0,2
	3		III.13-15	90	0,144	+0,493	0,329	0,456	0,427	0,413	0,3
	4		III.20-22	55	0,132	+0,589	0,336	0,458	0,430	0,416	0,3
	5		III.26	7	—	—	0,393	—	—	—	
1956	1	Mátra (4)	II.13-15	99	0,099	+1,005	0,136	0,177	0,166	0,161	0,1
	2		II.23-26	106	0,162	+1,495	0,220	0,338	0,305	0,289	0,2
	3		II.29-III.3	105	0,165	+1,342	0,266	0,408	0,369	0,350	0,3
	4		III.6-8	109	0,121	+1,064	0,306	0,420	0,390	0,376	0,3
	5		III.12-14	95	0,096	+0,347	0,285	0,356	0,340	0,333	0,3
	5/*a*	>650m		38	0,097	+1,085	0,273	0,355	0,333	0,323	0,3
	5/*b*	<650m		57	0,086	+0,013	0,293	0,352	0,340	0,334	0,3
	6		III.19-22	106	0,123	+0,129	0,329	0 427	0,407	0,397	0,3
	6/*a*	>650m		38	0,114	+0,725	0,301	0,398	0,375	0,364	0,3
	6/*b*	<650m		68	0,099	—0,007	0,346	0,426	0,410	0,402	0,3
	7		III.26-29	96	0,110	+0,012	0,330	0,416	0,399	0,390	0,3
	7/*a*	>650m		38	0,091	+0,639	0,320	0,401	0,382	0,373	0,3
	7/*b*	<650m		58	0,117	—0,332	0,336	0,418	0,404	0,396	0,3
	8		IV.2-5	32	0,083	—0,032	0,398	0,474	0,459	0,452	0,
	2	Tarján (5)	III.8-9	21	0,137	+0,091	0,310	0,412	0,391	0 381	0,3

Remarques : *n* = nombre des données
Cv = facteur de variation
Cs = facteur d'asymétrie

ssins versant des ruisseaux « Garadna », « Kövicses » et « Tarján » 1954-1956

densité kg/l										'différence'	
d'une probabilité de											
30	40	50	60	70	80	90	95	97	99	X_5—X_{95}	X_{10}—X_{90}
%										kg/l	
0,191	0,182	0,174	0,165	0,157	0,146	0,132	0,121	0,113	0,099	0,108	0,084
0,244	0,234	0,225	0,217	0,210	0,203	0,195	0,190	0,187	0,183	0,109	0,084
0,392	0,383	0,375	0,367	0,358	0,348	0,335	0,325	0,318	0,305	0,106	0,083
0,196	0,191	0,186	0,181	0,177	0,172	0,165	0,160	0,157	0,152	0,062	0,048
0,253	0,240	0,230	0,221	0,213	0,206	0,198	0,193	0,191	0,189	0,133	0,101
0,351	0,337	0,325	0 313	0,302	0,289	0,271	0,258	0,250	0,236	0,154	0,120
0,356	0,343	0,332	0,321	0,310	0,298	0,283	0,271	0,264	0,252	0,144	0,112
—	—	—	—	—	—	—	—	—	—	—	—
0,141	0,137	0,134	0,131	0,128	0,125	0,121	0,118	0,117	0,115	0,043	0,033
0,231	0,220	0,212	0,204	0,197	0,191	0,184	0,180	0,178	0,175	0,109	0,084
0,280	0,267	0,257	0,247	0,238	0,230	0,220	0,214	0,211	0,207	0,136	0,105
0,320	0,309	0,300	0,291	0,283	0,274	0,265	0,258	0,254	0,249	0,118	0,091
0,298	0,290	0,283	0,277	0,270	0,262	0,251	0,243	0,238	0,228	0,090	0,070
0,283	0,275	0,268	0,262	0,257	0,250	0,244	0,239	0,236	0,232	0,084	0,065
0,306	0,299	0,293	0,287	0,280	0,272	0,261	0,252	0,246	0,235	0,082	0,064
0,350	0,338	0,328	0,318	0,307	0,295	0,278	0,264	0,255	0,239	0,133	0,104
0,316	0,306	0,297	0,289	0,280	0,272	0,261	0,253	0,247	0,239	0,111	0,086
0,364	0,355	0,346	0,337	0,328	0,317	0,302	0,290	0,282	0,265	0,112	0,088
0,349	0,339	0,330	0,321	0,311	0,299	0,283	0,270	0,262	0,245	0,119	0,093
0,333	0,324	0,317	0,310	0,303	0,295	0,285	0,278	0,274	0,266	0,095	0,074
0,358	0,348	0,338	0,328	0,317	0,304	0,284	0,268	0,257	0,235	0,128	0,100
0,415	0,406	0,398	0,390	0,381	0,370	0,356	0,344	0,336	0,320	0,108	0,084
0,332	0,320	0,309	0,299	0,288	0,274	0,256	0,242	0,232	0,214	0,139	0,109

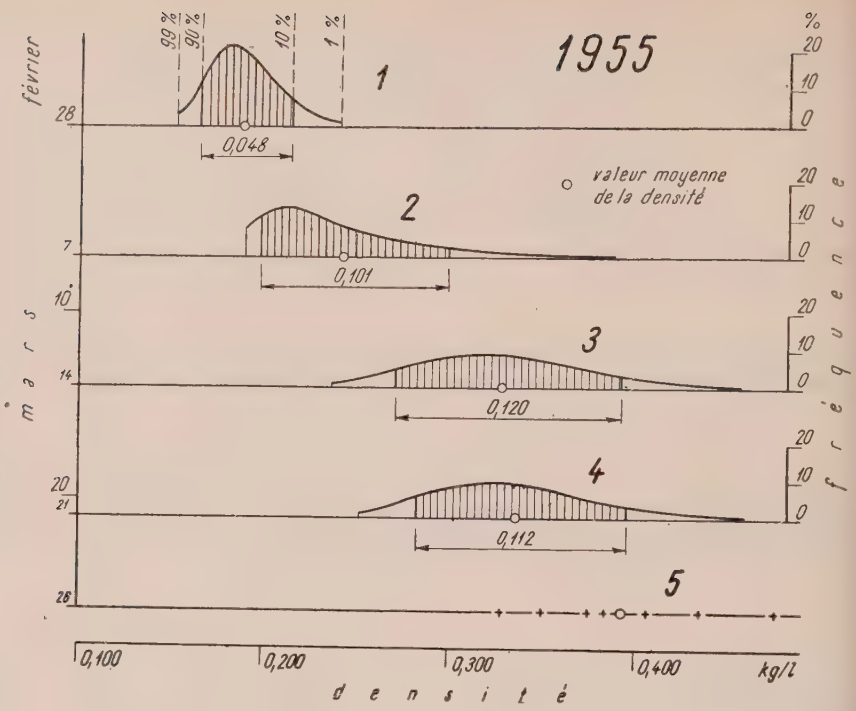

Fig. 6 — Étude de la répartition des valeurs de la densité dans l'espace et dans le temps. Mts. Mátra, 1955.

endroits abrités de la radiation, si elles avaient été recouvertes, par le vent, de loess, de la poussière et d'herbes sèches, au cours de la période de forte gelée. L'enneigement, composé à l'intérieur de deux à quatre couches congélées, à également une densité élevée [10].

Les effets du relief ont été mis en évidence surtout par les recherches effectuées dans les Bükk (7). Les résultats des mesures faites pendant les journées de 12 à 13 et de 16 à 28 février 1954, sont représentées sur la Figure 8, en coupes transversales verticales imaginées en direction sud-nord. On y fait figurer également les valeurs caractéristiques de l'épaisseur de l'enneigement et celles de son stock en eau. Dans l'intervalle de ces quinze jours de recherches, il n'y a pas eu d'importantes chutes de neige. L'augmentation de l'équivalent en eau est due surtout à la condensation qui se renouvelle par périodes de 24 heures. On peut constater, en consultant la Figure 8, que *la densité est extrêmement élevée aux pentes exposées au Sud, elle est moyenne aux plateaux et dans les fonds des vallées, et petite aux versants nord; elle décroit sensiblement avec l'augmentation de l'altitude.* Des mêmes recherches effectuées dans les Bükk, nous ont fourni un autre exemple encore de la répartition de la densité dans l'espace, sous l'effet des influences du relief. Nous le reproduisons sur la Figure 9 où l'on peut se rendre compte aisément des effets du relief. Au cours des recherches poursuivies dans les monts Mátra (4), l'influence du relief s'est manifestée de la manière la plus complète dans la vallée asymétrique du Kövicses. [9]

Avant d'examiner de plus près la diminution accentuée de la densité par rapport

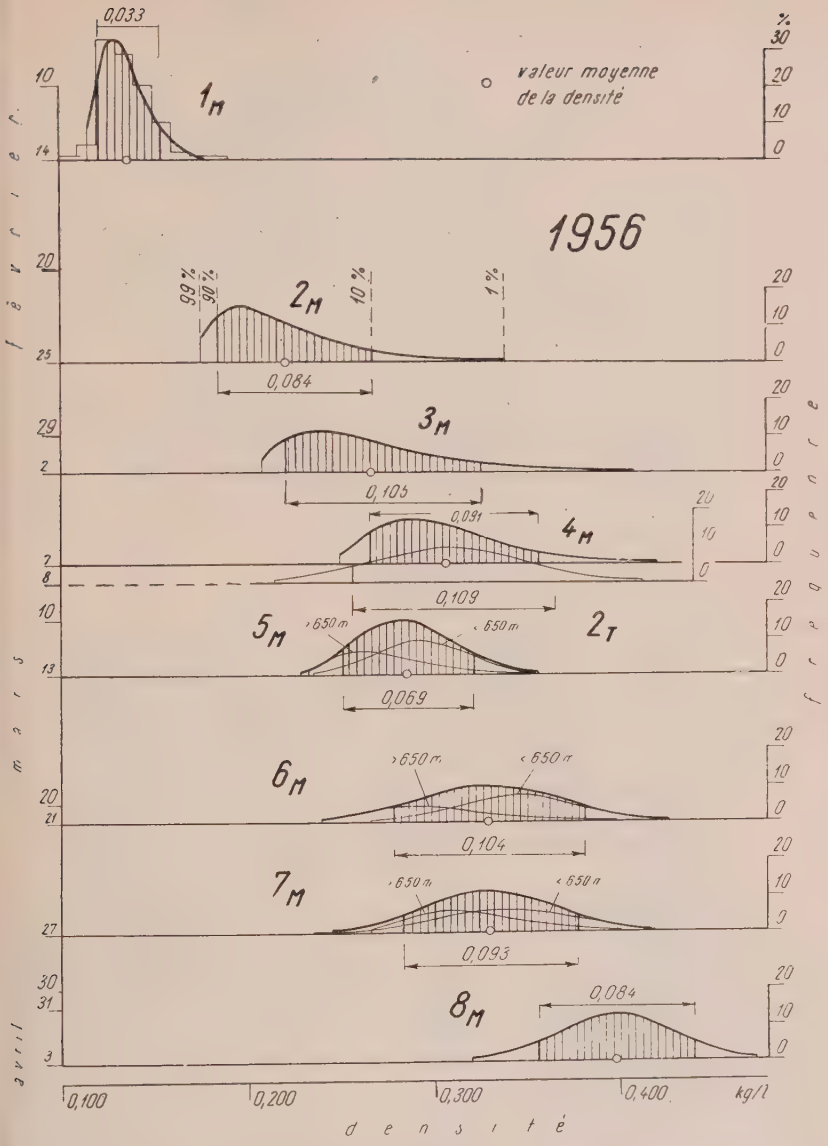

Fig. 7 — Étude de la répartition des valeurs de la densité dans l'espace et dans le temps. Mts. Mátra et le bassin versant du ruisseau du Tarján, 1956.

à l'augmentation de l'altitude, nous traiterons d'abord du changement de la durée de l'enneigement qui accompagne l'augmentation de l'altitude.

La durée de l'enneigement, selon les valeurs moyennes des années 1929/30-1943/44, *s'accroît*, par 100 m, de 6 à 10 jours dans la Transdanubie et de 20 à 30 jours dans

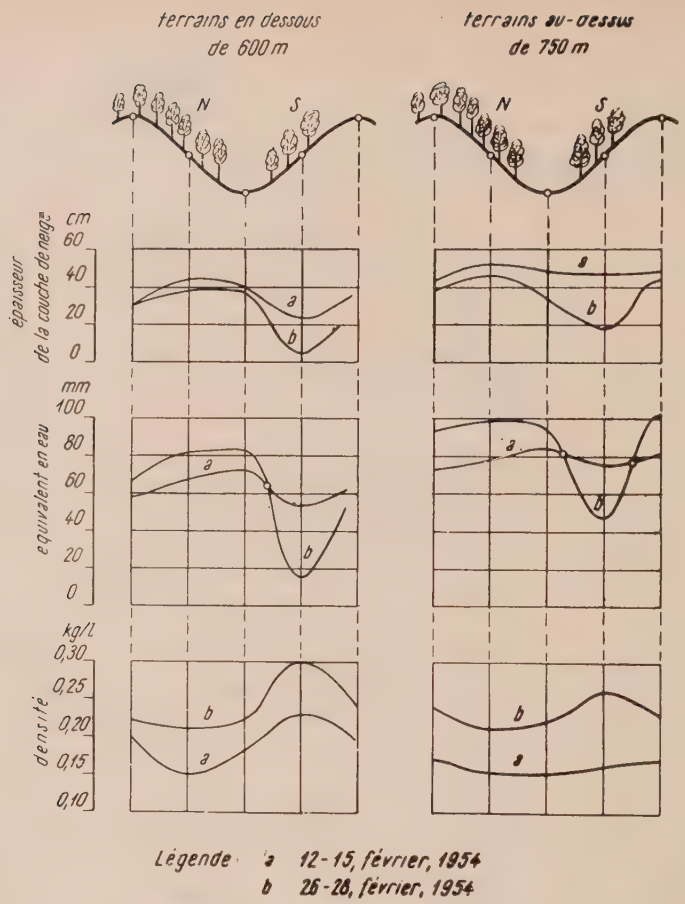

Fig. 8 — L'influence des facteurs de relief. Mts. Bükk, 1954.

la région montagneuse nord, jusqu'à 400 m d'altitude à peu près (Figure 10). Dans les régions montagneuses qui peuvent être considérées comme les plus élevées, par rapport aux conditions géographiques de la Hongrie, c'est-à-dire au-dessus de 400 m environs, elle ne s'accroît plus que de 10 à 18 jours, par 100 m de différence d'altitude; l'augmentation est donc beaucoup plus faible ici que n'est l'augmentation des valeurs totalisées dans la partie Est du pays (l'Alföld et la région montagneuse), dont l'altitude reste en dessous de 400 m. Pendant les quatre hivers compris dans l'intervalle de 1952/53-1955/56, années où nous avons effectué la plupart de nos recherches, la durée moyenne, par rapport au pays, était de 20 jours par 100 m, en allant jusqu'à 200 m; elle s'accroissait de 15 jours pour les altitudes comprises entre 200 et 600 m, et de 10 jours par 100 m pour les altitudes dépassant les 600 m. On peut constater que les données de cette espace de temps, groupées par niveaux d'altitude, s'accordent très bien avec la moyenne des relevées d'un grand nombre d'années. Nous représentons dans le Tableau IV, l'accroissance du nombre des jours d'enneigement par rapport à l'augmentation de l'altitude, répartie selon les régions du pays. Comme

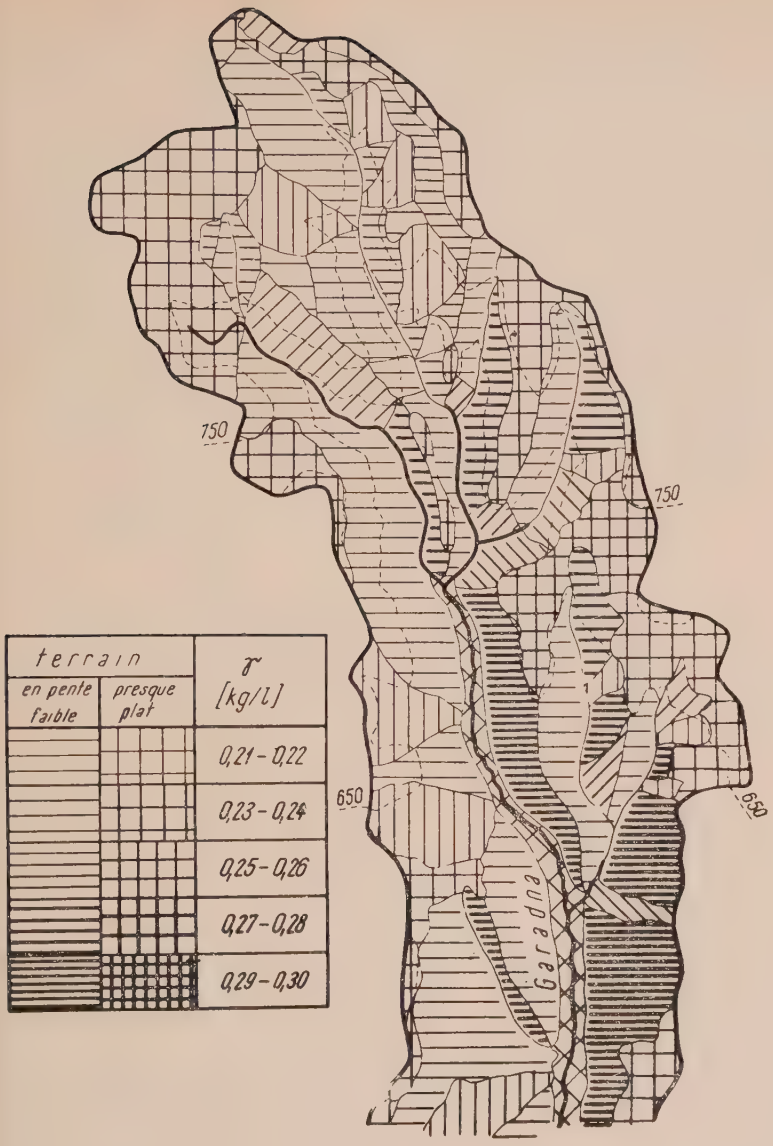

Fig. 9 — L'influence des facteurs de relief. Mts. Bükk, 26-28, II, 1954.

on le voit par les résultats, *la durée de l'enneigement était un peut plus grande à l'Alföld, dans les plaines en dessous de* 100 *m, que dans le pays de collines, entre* 100 *et* 200 *m.* Cette inversion, de même que celle que nous avons constaté dans la moyenne d'un grande nombre d'années, remontent aux influences de la structure du sol. Dans la grande plaine hongroise, en effet, les sols des terrains bas sont. généralement,

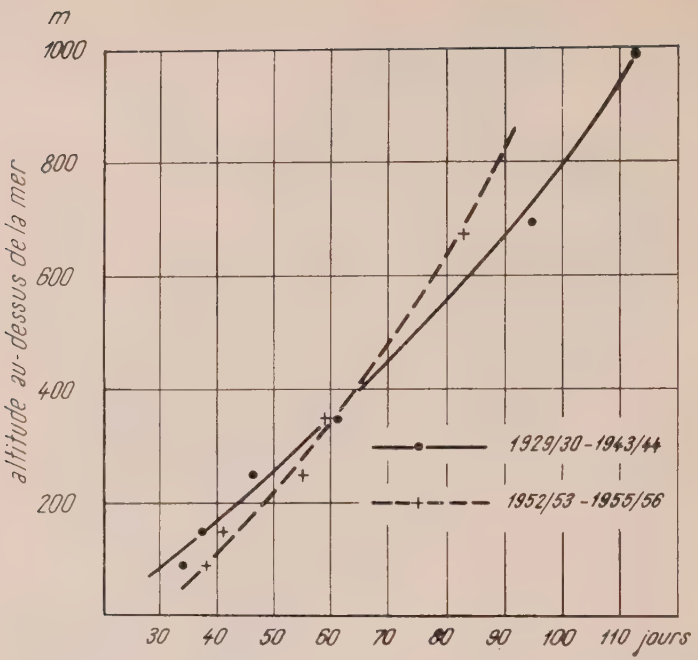

Fig. 10 — Le changement du nombre des jours d'enneigement avec l'altitude, en Hongrie.

TABLEAU IV

Augmentation du nombre des jours d'enneigement 1952/53-1953/54

Régions naturelles du pays	Altitude en m, au-dessus du niveau de la mer Adriatique				
	100	100-200	200-300	300-400	400
	nombre des jours d'enneigement				
Transdanubie	37	44	49	60	—
Alföld (grande plaine hongroise)	39	38	62	—	—
Montagnes du Nord	37	45	53	59	83*

*à l'altitude de 672 m

froids, compacts et limoneux, tandis que les terrains plus hauts — entre le Danube et la Tisza surtout — se composent de sols sableux, chauds, légers et friables; il est donc naturel que sur les premiers, l'enneigement ne garde plus longtemps que sur les derniers. C'est pour la même raison que, dans l'Alföld, le manteau de neige recouvre pendant un temps plus long les terrains dont l'altitude n'atteint pas les 100 m, que dans la Transdanubie où le sol est médiocrement compact. Il en est de même pour les terrains ayant une altitude de 100 à 200 m dans l'Alföld, par suite de la structure du sol, ils restent recouverts de neige plus longtemps que dans la Transdanubie, à la même altitude. La haute valeur extraordinaire de l'altitude de 200 à 300 m de la grande plaine hongroise semble confirmer la thèse de géographie physique suivant laquelle la durée de la vie de l'enneigement des montagnes massives serait moins longue que celle des montagnes isolées [1]. En résumé, nous devons insister sur le fait que — les niveaux d'altitude de 100-200 m de l'Alföld exceptés — *l'augmentation de la durée de l'enneigement accompagne toujours l'accroissance de l'altitude*, et cela paraît avoir presque le caractère d'une loi.

Cette augmentation entraîne naturellement la diminution de la densité, fait que nous avons déjà démontré sur la Figure 8, en y reproduisant les valeurs de la densité par rapport à deux coupes imaginaires d'une hauteur différente. Sur la Figure 11

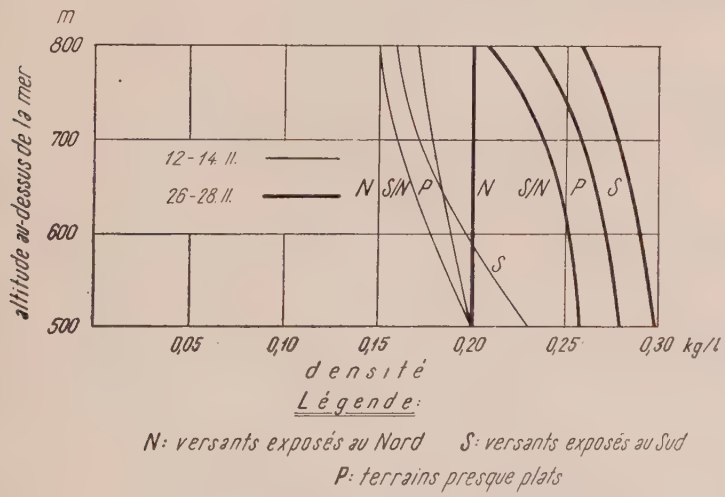

Fig. 11 — L'influence des facteurs de relief et de l'altitude. Mts. Bükk, 1954.

nous représentons ces données d'une manière plus directe, en fonction de l'altitude. On voit que, si l'on excepte l'enneigement du versant nord abrité, ne subissant aucun changement, la densité diminue partout. [9]

Les recherches effectuées dans les monts Mátra (4) ont fourni également la preuve de la thèse relative à la diminution de la densité par rapport à l'augmentation de l'altitude. Il suffira d'en citer ici un exemple : les mesures effectuées, au cours de la quatrième semaine, à des altitudes différentes, ont donné pour la densité moyenne, les valeurs suivantes :

> 850 m $\gamma = 0{,}287$ kg/l
500-850 m $= 0{,}298$ »
350-500 m $= 0{,}331$ »
< 350 m $= 0{,}378$ »

Nous avons déjà établi l'influence de l'altitude en étudiant la répartition des relevés de mesure recueillies dans les monts Mátra (4). Les résultats qu'on avait obtenu au cours des cinquième, sixi.ème et septième semaines en 1956, ont été classés en deux groupes (zone au-dessus et zone en dessous de 650 m) et controlés par l'approximation effectuée à l'aide de la courbe de répartition de la formule III de Pearson. Le Tableau III et la Figure 5 permettent de se rendre compte des différences qui se produisent à la suite de l'influence de l'altitude : le maximum de la fonction de répartition se déplace et la différence des densités de la probabilité de 10 % et de la probabilité de 90 % diminue.

Les influences de la végétation ont été vérifiées aux cours des recherches poursuivies dans les monts Bükk (7) et les montagnes de Buda (3). La densité a atteint la valeur la plus élevée sur les prés et pâturages plus ou moins étendus, entourés de forêts. Elle a été la plus petite dans les forêts d'âge moyen, et avait une valeur intermédiaire dans les vieilles et les jeunes forêts (Tableau V). Il ressort de là que l'effet climatique des arbres dont l'ombre recouvre le sol complètement en été, reste valable même en hiver. On peut constater. en général, que le changement de la densité de l'enneigement est beaucoup moins considérable dans le cas de la variation des groupes de plantes que dans celui de l'altitude. [9]

TABLEAU V

Influence de la végétation sur l'économie d'eau de l'enneigemnet
(bassin versant du ruisseau du Garadna)

végétation	de 12 à 15 II. 1954				de 26 à 28 II. 1954			
	n	v	b	γ	n	v	b	γ
		cm	mm	kg/l		cm	mm	kg/l
pré, pâturage	18	44,4	73,5	0,165	25	38,6	92,8	0,241
vieille forêt	39	43.1	73,0	0,169	71	37,2	83,0	0,223
forêt	25	41,3	63,5	0,154	27	36,3	79,2	0,218
jeune forêt	19	42,1	70,2	0,166	29	38,0	85,2	0,224

Remarques : v = l'épaisseur de la couche de neige,
h = l'équivalent en eaux de la couche de neige,
γ = la densité de la couche de neige,
n = nombre des prélèvements d'échantillons.

Le changement de la densité dans le temps

La densité peut varier dans le temps, sous les influences climatiques, de la manière la plus diverse. Le motif principal des changements dans le temps, c'est *le changement périodique de* 24 *heures* produit, le plus souvent, par les alternatives de gel et de dégel du jour et de la nuit. C'est de ces changements quotidients, se déroulant en peu de temps, que se compose *le processus du changement complet* embrassant toute la durée de la vie de l'enneigement. Dans ce qui suit, nous exposons les résultats de nos recherches par rapport au processus complet d'accumulation et de fonte de l'enneigement.

Aux conditions climatiques de la Hongrie, dans une année d'enneigement durable, *le changement de la densité dans le temps, s'effectue d'une façon assez identique sur les territoires les plus éloignés du pays*, jusqu'au commencement de la fonte de neige. Sur la Figure 12, nous présentons les changements de la densité observés, pendant les mois de janvier et de février 1954, dans les montagnes de Buda, au jardin de

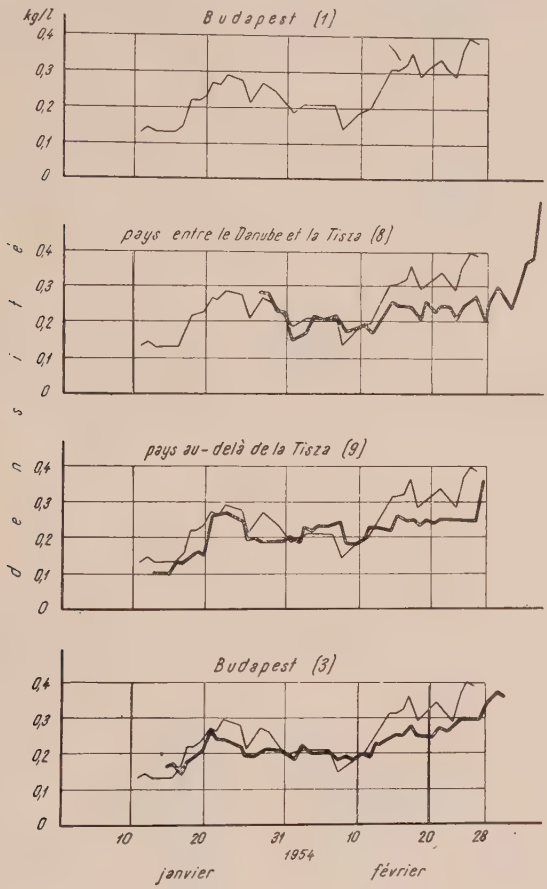

Fig. 12 — Le changement de la densité dans le temps. I-II, 1954.

l'Observatoire astronomique (3), entre le Danube et la Tisza, à la station « Komlósi » (8) et au-delà de la Tisza, à Taskony (9), le tout par rapport aux observations recueillies dans le jardin de l'Université des Sciences Techniques de Budapest (1). Il faut naturellement insister sur le fait que, pendant l'hiver en question, les conditions de l'économie thermique et de l'économie d'eau avaient été presque les mêmes dans tout le pays. [10]

Il est à remarquer que dans *le changement de la densité*, on peut observer des phénomènes qui permettent *de faire des estimations*. Nous en mentionnerons deux : le rapport entre le changement de la densité et la fonte, et les rapports entre le changement de le densité et l'épaisseur de l'enneigement. ([7—11]

On peut constater, tout d'abord, qu'il est possible de déterminer *la valeur limite inférieure* de la densité (γ_i). Avant qu'elle apparaisse, chaque goutte de l'eau de fonte reste dans l'enneigement en tant qu'eau capillaire et eau se tranformant en glace. Il est possible de déterminer *la valeur limite supérieure* (γ_s) aussi. Au moment où elle est atteinte, la fonte est déjà commencée, l'enneigement perd de l'eau qui suinte à travers la couche de neige et s'écoule. La valeur limite inférieure, dans les conditions de Hongrie, est à γ_i = 0,22-0,25 kg/l, la valeur limite supérieure à γ_s = 0,35-0,40. Si la densité est entre les deux valeurs limites, l'eau de fonte reste dans la neige ou découle, selon la situation momentanée de l'économie d'énergie thermique et de l'économie d'eau. [10]

L'observation nous a amené à constater qu'au cours de la fonte se produisant *entre les limites inférieures et supérieures*, il existe une corrélation assez étroite entre la diminution de l'épaisseur de l'enneigement et l'augmentation de la densité. *A une diminution de* 1 *cm de l'enneigement, correspond approximativement une augmentation de* 0.01 *de la densité.* [10]

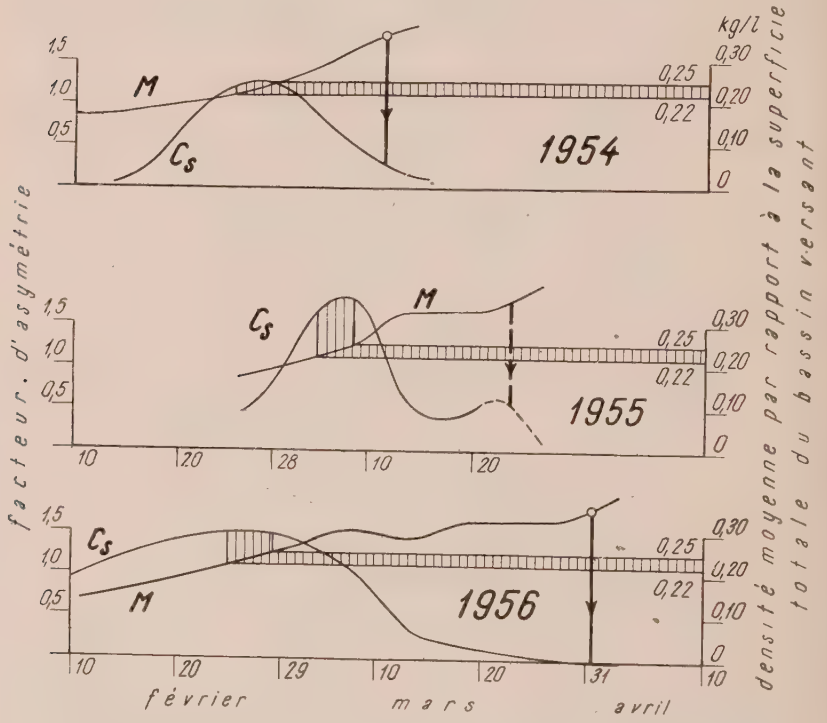

Fig. 13 — Le rapport entre la valeur moyenne de la densité (M) et le facteur de l'asymétrie (C).

A partir de ces observations et constatations, nous avons élaboré *une méthode pratique* servant à l'estimation du changement de la densité. Nous en présentons les principes de base et les rapports fondamentaux dans le Tableau VI. [11]

Par rapport aux changements se produisant dans le temps, on peut découvrir

des corrélations étroites aussi entre les résultats obtenus dans l'étude du bassin versant entier (Tableau III, Figures 5-7).

Le coefficient d'asymétrie (Cs) caractérisant la répartition des résultats des observations recueillies sur un bassin versant, *est une valeur relativement peu élevée* (Tableau III. 1954 : +0,071). Dans le cas d'une transmission de chaleur minime à l'enneigement, le coefficient d'asymétrie s'*accroît brusquement* (Tableau III, en 1954 : +1,252, en 1955 : +0,622 et après : +1.584, en 1956 : +1,005, et après : +1,495). *La valeur maximum* du coefficient d'asymétrie apparaît à l'enneigement dont la densité moyenne est $\gamma_i = 0{,}22\text{-}0{,}25$ (Figure 13).

TABLEAU VI

Formules servant à la détermination de la densité de la neige

domaine de la densité	calcul de la densité (kg/l) si le jour « $j-1$ »				
	il n'y avait pas de précipitation	il y avait beaucoup de pluie	il y avait peu de pluie	il neigeait	
$\gamma < \gamma_i$	$\gamma_j = \frac{\sum h}{10\ v_j}$				(1)
$\gamma_i < \gamma < \gamma_s$	$\gamma_j = \gamma_{j-1} + (v_{j-1} - v_j)\ 0{,}01$	(2)	$\gamma_j = \frac{H_{j-1} + h_{j-1}}{10\ v_j}$		(3)
γ_a atteint γ_s	$\gamma_j = \gamma_s$	(4)		où $H_{j-1} = 10\gamma_{j-1}\ v_{i-1}$	

Remarques : γ_i et γ_s *v.* le texte,
v l'épaisseur de l'enneigement en cm,
H l'équivalent en eau de l'enneigement en mm,
h neige en mm ou pluie en quantité minime n'entraînant pas avec elle l'affaissement de l'enneigement, aussi en mm

Dans ce cas la répartition des valeurs de densité est particulièrement intéressante. En 1956 (23-26 II.), par exemple, la densité a été, dans les 77 % des cas, entre $\gamma = 0{,}18$ et $\gamma = 0{,}25$. Il y a eu à peine une ou deux valeurs (5 %) qui restaient en dessous de 0,18 .Les 18 % des cas dépassaient les 0,25, conformément à l'asymétrie. Ce résultat surprenant peut nous amener à faire plusieurs constatations. Notamment :

L'enneigement composé de neige nouvellement tombée et dont la densité n'atteint pas le 0,18 *ne peut être considéré du point de vue de sa structure, comme durable*; au moindre effet de la chaleur, il perd sa structure originale.

L'enneigement dont la densité est comprise entre 0,18 et 0,25 *est d'une structure relativement plus stable*; à la première phase de la fonte, une grande majorité des cas se trouve dans ce domaine. C'est dans ce domaine, que nous avons fixé la valeur extrême inférieure aussi, caractérisant la fonte (γ_i). Le choix fut confirmé par le

calcul des probabilités exposées plus haut, puisque l'enneigement ayant la situation la plus stable et n'abandonnant rien de son eau se trouve, en effet, également dans ce domaine. Au-dessus de la densité de 0,25, en conformité avec les différents effets de chaleur auxquels l'enneigement avait été exposé (des « recettes » de la radiation plus importante aux versants orientés vers le Sud), nous avons observé des valeurs de densité allant de 0,25 à 0,35. Le fait que le cercle des relevés s'élargit, est des plus naturels, puisque la valeur de densité γ_s = 0,35-0,40, designant la limite supérieure de la fonte, est encore éloignée.

Dans les phases ultérieures de la fonte, *le coefficient d'asymétrie ne fait que diminuer*; sa régression correspond au changement relativement lent qui s'opère dans la structure de l'enneigement, entre les densités de γ = 0,25 et γ = 0,35, donc les effets thermiques, suscitant la fonte, peuvent se développer librement.

A la dernière étape de la fonte, le *coefficient d'asymétrie peut devenir négatif* (observation en 1956, à la huitième semaine des recherches : —0,032), conformément à la contrainte exercée par la valeur limite supérieure (que la densité ne peut pas dépasser de beaucoup). Le coefficient d'asymétrie négatif *justifie la valeur limite supérieure* établie à la base du calcul des probabilités.

La valeur de la densité peut varier, pendant tout le processus de l'accumulation et de dégel, sur une très large échelle, conformément à la situation climatique momentanée. Mais, à la naissance et à la disparition de l'enneigement, on peut observer des lois qui paraissent être rigoureusement déterminées. Au moment où la neige tombe, et peu après aussi, on ne peut constater que des différences minimes (la différence entre les valeurs de probabilités de 10 %, et de 90 % fut, 0,048 en 1954, à la première semaine des recherches, et seulement 0,033 en 1956, Tableau III), *ce qui prouve la construction de structure uniforme de l'enneigement*. Lors de sa disparition définitive, à partir du moment où s'établit γ_s = 0,35-0,40, la variation des valeurs de la densité diminue ou, tout au moins, ne s'accroît plus. Dans ce fait, on peut voir *une nouvelle justification de la valeur extrême supérieure*.

* * *

La connaissance des lois relatives aux changements de la densité de l'enneigement permet d'évaluer, d'une façon plus précise, son équivalent en eau, et de déterminer la quantité de la neige fondue. Grâce à elle, il sera possible de développer d'avantage les recherches sur l'économie d'énergie thermique et de l'économie d'eau. Dans ces domaines, on a pu déjà élaborer et introduire certains procédés, et contribuer à la solution de plusieurs problèmes hydrologiques [9—11]. Ces lois ont été établies dans les conditions climatiques de la Hongrie et, en tout rigueur, elles ne sont valables que dans les cadres de ce pays. Nous espérons pourtant qu'elles pourraient être utilisées, dans des conditions climatiques analogues, même dans d'autres pays.

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FOREST-COVER EFFECTS ON SNOWPACK ACCUMULATION AND SNOW MELTING IN RELATION TO METEOROLOGICAL CONDITIONS

V. V. RAKHMANOV

Summary

The problem of forest influence on the accumulation and thawing of snow is a part of the problem of its role in water preservation. A correct understanding of this influence is necessary to study water regime of rivers and plan forest reclamation.

The study of snow measuring data shows that the ratio of snow reserves accumulating by the end of winter in the forest and in the field depends not only on the character of forest plantations (composition, thickness, etc.) but also on meteorological conditions of winter.

It is considered that water reserves in the snow cover of forests by the end of winter are usually less than in the field because some quantity of snow is kept by tree tops and evaporates.

The data of parallel snow surveys at 89 hydrometeorological stations in the central European part of the USSR and partially in Western Siberia for the 1947-1954 period lead to the conclusion that in more than 90 per cent of all cases snow reserves in the forest (in terms of water) by the end of winter exceed those in the field. An average difference between them is 18 per cent.

But this difference is not constant; it changes from 5—10 to 20—25 per cent in according to meteorological conditions during the period of snow accumulation.

The analysis of synoptical processes over the European part of the USSR shows that if winter is accompanied by cyclonic activity and frequent stormy weather changes, the snowreserves in the forest exceed those in the field still more. On the contrary, if there is a lingering weather of an anticyclonic nature, this difference decreases.

The variation of a relation between forest and field snow reserves with allowance for temperature and precipitation variations in winter time is well effected even from snow survey at separate stations, both in the West and in the East of the European territory of the USSR. From joint diagrams of snow and average diurnal (and maximum) positive temperatures of the air it can be seen that snow reserves in the forest relatively increase with the arrival of warmer weather.

The comparison of air temperature variations and changes of water contents in snow permits to find the dependence between the difference of forest and field snow reserves and the sum of positive average daily temperatures for the period of snow accumulation. This difference between snow reserves rises with an increase of the sum of the temperatures.

Such relations can be utilized for approximate estimation of forest snow reserves by snow reserves in an open terrain with allowance for meteorological conditions in the previous period of snow accumulation. An increase of the ratio between forest and field snow reserves is explained by a longer detention of cold air in forests which hampers the snow thawing and decreases evaporation from the snow surface.

The same causes delay the spring thawing of snow in the forest where even in cloudy weather the snow thaws slower than in the field. But due to the weather in the period of thawing the time difference of snow thawing in the field and in the forest can vary from one or two days to several weeks. Consequently, not only the kind of plantations but also weather in winter determine the controlling influence of forests on a run-off of rivers during high-floods. It is best seen in mutable weather when warm air masses change for cold ones and vice versa. The snow in the forest does not melt for a long time in clear anticyclonic weather when the air is getting warm slowly.

The account of these and other peculiarities of weather in the period of snow thawing provides for a forecast of a general development of high-floods in the rivers in forest regions.

The question of forest influence on the formation of snow cover and snow melting is an important problem of human activity, of the water conservation. The correct understanding of these effects is necassary for planning different measures in forestry, agriculture and for improving hydrological regime of rivers.

For a long time snow was considered to accumulate more in forests than in open areas, including fields. This opinion was based on the notion of snow cover height which in forests is usually known to be higher than in fields.

However, when snow-surveys have been begun, including the measurement of height, density in the open areas and in forests, it was discovered that the water content of snow in forests is sometimes smaller than that of the open areas. In the 80-s of the last century the first to make parallel (synchronized) snow-surveys near Moscow was Tursky. His surveys showed that by march there accumulates more snow in the open areas than in the forests under the trees; in dense forests there accumulates less snow than in thin forests and in coniferous forests — less than in deciduous forests. Tursky explained it by the fact that during snowfall in forests much snow remains on the branches of the trees. The results of his investigations were highly appreciated by Voeykov, a well-known climatologyst and an initiator of the snow cover study.

Subsequently snow-surveys came to be carried on in other countries where snow-cover plays a substantial part of the river water balance. However, the results of these snow-surveys were contradictory. While some of them confirmed the opinion of a smaller snow accumulation in forests the others, on the contrary, showed that in forests by the end of winter there accumulates more snow than in fields. Finally, there have been obtained a great number of data, according to which snowpack accumulation in the open areas are larger than in coniferous forests and less than in deciduous and mixed forests.

One of the reasons of these contradictions lays, probably, in different methods of snow-surveys. However, to our opinion this is not the main reason. The thing is that the above conclusions were drawn from snow-survey data which were carried on not only in different afforestations, various in composition, volume, age and other characteristics, but also in different places and in different years, that is, in different meteorological conditions.

Many scientists (3, 5, 10, 12) who studied snow and snow cover have already paid an attention to the influence of meteorological conditions on the snow accumulation of forests. However, the regularity of this accumumlation under various meteorological conditions is still poorly known.

The paper gives the results of investigations of snow-survey data which were carried in on forests and fields in the period of 8-10 years at 89 hydrological stations of middle belt of the European territory of the USSR and some stations of the Western Siberia. The investigation does not include the data of those stations where the results of snow-surveys were distorted by wrong location of snow-surveying plots (in the border of forests, in the places of intense snow blowing).

The forest-cover effects on snowpack accumulation were revealed by comparison of water content in forest snow-cover with that in the fields. For this purpose the term of *forest snow-storage coefficient* K was introduced meaning the relation of snow water storage in forests to water storage of adjacent field or other non-afforested area. When the water storage of snow in forests and in fields is the same (for short, they will by called snow-storage), the snow-storage coefficient is equal to 1.

Table I gives the mean snow-storage coefficients in forests which were calculated for the dates of maximum snowpack accumulation for the two regions: Western region, including Byelorussian SSR and Western areas of the RSFSR, and Eastern region, including the Ural and adjacent areas of the European territory of the USSR, and partly of the Western Syberia. The Western region is characterized by unsteady winter with frequent thaw, which causes sharp changes of snow-storage. On the contrary, the Eastern region has continental climate with cold winter and stable snow cover.

As it is seen from the table, mean coefficients of snow-storage in forests in the

Western as well as in the Eastern regions are approximately equal, in spite of different climatic conditions. It may be considered that in winter in the forests of the Western and Eastern parts of the European territory of the USSR the snow accumulation is 18 per cent more than in fields and other open areas. The number of occurrences when the snow storage of forests was less than that of fields amounts in the Western area to 12.7 per cent, and in the Eastern area — to 4.5 per cent. Attention must be paid to the slight unsteadiness of snow-storage coefficients that is confirmed by small variation of coefficients in the both regions.

TABLE I

Regions	Number of points	Number of year stations	Snow-storage coefficient			Variation coefficient of snow-storage coefficient	Number of points where $K<1$	
			average	minim.	maxim.		number	per cent
Western	63	281	1.18	0.88	1.63	0.18	8	12.7
Eastern	26	136	1.18	0.95	1.41	0.10	1	4.5

So the numerous data of snow-surveys show that in winter the forests favour the accumulation of great amount of water in snow in comparison with open areas in the Western areas with unsteady snow cover, as well as in the Eastern areas, where snow cover is less subjected to thaws.

The character of afforestation is known to influence greatly the snow accumulation. From the same snow-surveys it is ascertained that in the deciduous forests, as well as in the mixed forests with a great number of deciduous species the average coefficient of snowstorage in the Western region amounts to 1,35, and in the Eastern— to 1.27. In the coniferous forests it is correspondingly equal to 1.12 and 1.08. It means that in winter period the volume of water in snow in the deciduous forest averages 27-35 per cent more that in the open plots. Undoubtedly, some thick forests, firgroves, for instance. may contain less snow than field .The given ciphers refer mainly to all the forests, covering river basins because the points of snow-surveys are set in the forests, typical for the location of hydrometeorological stations.

The obtained results can be applied to other parts of the European territory of the country and it is proved by the investiagations of the snow cover in the Northern region. (7).

Hence, it follows that while calculating the run-off volume on basin territory it must be taken into consideration not only the whole area of forests in the catchment area, but also its distribution, according to the kind of forests.

However, the snow accumulation in the forrests depends not only on the character of afforestation, but, as is already mentioned, also on the meteorological conditions, in other words — on the weather course during winter season. The calculation of the means snow storage coefficients for all points of each region, shows that they are not constant, but vary from year to year, remaining, however, more than 1. (Table II). This means that in our meteorological conditions according to the calculation of many years, as well as separate years, there accumulates more snow in all the forests than in the fields.

TABLE II

Mean snow storage cofficient variation in the forests with each year:

Regions	Years									Means
	1945 1946	1946 1947	1947 1948	1948 1949	1949 1950	1950 1951	1951 1952	1952 1953	1953 1954	
Western	—	—	1.23	1.16	1.12	1.18	1.17	1.21	1.23	1.18
Eastern	1.16	1.08	1.22	1.14	1.15	1.15	1.16	1.18	1.16	1.18

The table II shows that the snow-storage coefficient in forest varies with each year. In some years it equals 1.08-1.12, in others it grows to 1.22-1.23. While, sometimes, the snow-storage in forests exceeds the snow-storages in fields only by 10 percent, in other years it almost reaches 25 per cent. In the latter case the role of forests in the formation of water regime considerably increases.

The analysis of the synoptic processes in the winter periods of 1947-1954 shows that the snow-storage coefficients reach the greatest value in winters characterized by the most active cyclonic activity and, consequently, by the frequent changes of weather in the period of snow accumulation. Such processes prevailed over the whole European territory of the USSR during the winters of 1947-48, 1950-1951, 1951-1952. In the winters of 1946-47, 1952-53 and 1953-54 the predominance of cyclonic activity was observed in the Western part of the country, while cold weather of anticyclonic character prevailed over the eastern part. The winters of 1948-49 and 1949-50 were characterized by weakened cyclonic activity almost over the whole European territory of the USSR, and snow-storage coefficients for these years were not large in the both regions as well as the centre of the European territory of the USSR which here is not considered.

Hence, during the winters with active cyclonic activity, when frequent changes of weather are observed, the fall of temperature is followed by its rise, leading in the western regions to thaws; snow storage in forests appear to be relatively greater in comparison with the open areas, than during the winters with relatively stable cold weather.

The relationship of snowpack accumulation in forests and the meteorological conditions in winter can be well traced from the snow-surveying data of separate stations. The fig. I gives the combined graphs of snow-storage variation in mm of water layer in fields and in forests and also of the mean diurnal air temperature and precipitation variation in 1952-53 at the hydrometeorological stations of Vassilevichi, Byelorussian SSR, and Shumikha, Kurgan region.

It is seen from the graphs that at the beginning of snow cover settling the snow storage in forests enlarges quicker than in the fields. Maximum winter correlation of forest and field snow-storages is usually fixed by the end of the winter when the fields accumulate the greatest volume of snow. In the period of spring snow-melting the correlation of snow storages enlarges still more, but it will be considered later.

Comparing the course of snow accumulation with the course of air temperature it is seen that the difference between the volume of snow storage in forests and fields grows mainly in the periods when the mean diurnal temperatures have positive meanings grows mainly in the periods when the mean diurnal temperatures have positive meanings or when it approaches 0^0, which is characteristic to the Eastern regions.

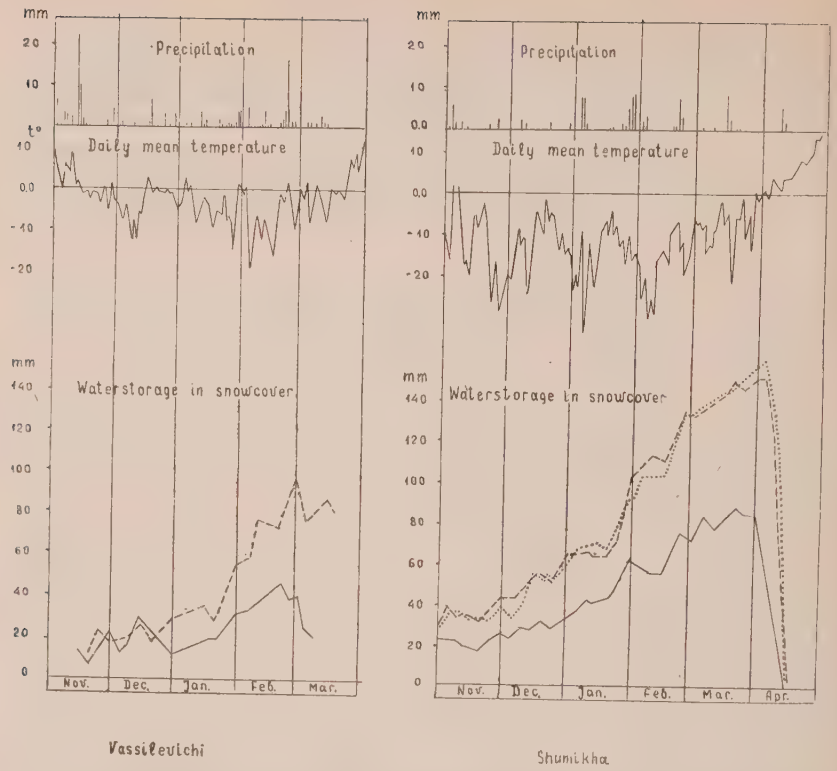

Fig. 1 — The graphs of water storage variations (in mm) in snow cover in forests and in fields, of mean diurnal air temperature and of precipitation in 1952/53, on the hydrometeorological stations of Vassilevichi (Byelorussian SSR) and Shumikha (Kurganski region). Water storage in snow cover: ——— in fields, — — — in forests, ... on the forest clearings. Ciphers of the temperature graph — the temperature at 1,00 p.m.

This means that in clear sunny weather, frequent in these regions, the daily temperatures become positive. When the air temperature rises the snowpack accumulation in fields decreases or increases slower than in forests, which is seen during abundant snowfalls accompanied by snow melting. Finally, by the end of the winter the snowpack accumulation in forests appear to be greater than in fields or in open areas. This is characteristic of the Western, as well as the Eastern, regions. Thus, separate station can well show the relationship of snow-storage correlation variation of forests and fields with the variation of the winter meteorological conditions, that is, with the rise of air temperature. The more frequent the rises of air temperature are, the more the snow storages in forests increase, which is revealed in the growth of snow-storage coefficients. These conclusions, drawn from the data of separate stations, fully agree with the above conclusions, drawn from the analysis of mean snow-storage coefficients variations in regions.

But if there is a relationship between relative snow storage increase and the rising of air temperature there must inevitably exist a relationship between the snow-storage coefficients or the difference of absolute snow-storage volume in snow, on

son hand, and the values of positive temperatures or their sums for the period of snow accumulation on the the other hand. In reality it is this relationship that is observed in nature. The most simple kind of this relationship is given on the fig. 2,

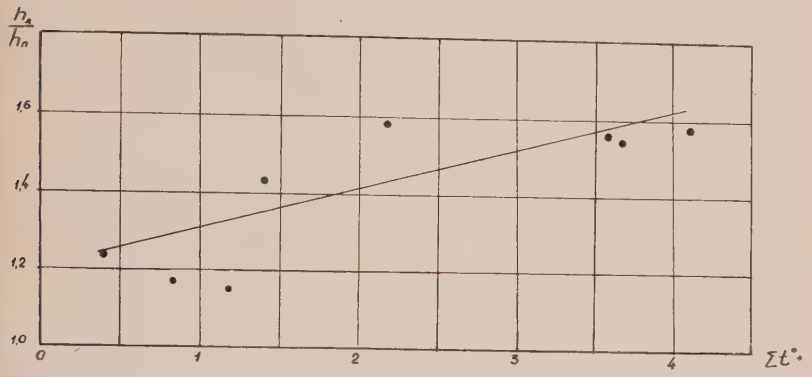

Fig. 2 — The graph of relationship between water-storage coefficients in forests ($\frac{hV}{h\pi}$) and the sum of mean diurnal positive air temperatures for the period of snow accumulation ($\Sigma t^0 +$) on the hydrometeorological station of Vissilevichi (Byelorussian SSR).

where a graph of relationship between snow-storage coefficients in forests and the sum of mean diurnal positive air temperatures at the hydrometeorological station of Vassilevichi (Byelorussian SSR) is shown. At this station the plots of snow surveys are in fields and in beech-hornbeam forests of mean thickness. The graph clearly shows that with the increasing of the sum of mean diurnal positive air temperatures in the period of snow accumulation, the snow-storage coefficient increases. In stable cold winter, when positive temperature sums are very small, snow storage in forests approache the snow storage in fields, and snow-storage coefficient decreases to the values close to 1.

Analogous relationship can be traced at many stations also between the snow-storage coefficients and the sums of daily positive air temperatures, for instance, the temperatures at 13 hours. Such relationships allow roughly to determine the snow-storage correlation in the forests and in the fields by the end of winter, basing on the sum of positive temperatures for the previous winter period. However, these relations are true only under definite conditions. In the time of intense thaws, when snow in forests melts completely away, they cannot be used: snow-storage coefficients increase to infinity. Therefore, it would be more correct to use the difference of absolute values of snow storage in forests and fields instead of the coefficients; at that there are the same relationships that are shown on the fig. 2, however they can be applied for all conditions of snowpack accumulation.

In order to understand the dynamics of the formation of snowpack accumulation in forests, relatively larger than that in fields, let us assume that snow accumulation is taking place under stable cold weather with negative temperatures, much lower than 0^0. If the snow cover is not disturbed by snow transmissions (snowstorms), there must accumulate by such weather less snow in forests than in fields, because some portion of snow stays on the tree crowns and then some of it evaporates. In such weather the volume of condensed or «horizontal» precipitation is small. The snow keeping influence of the tree crowns is mostly seen in thick orests, especially in fir-groves ([9]).

However, snow cover formation usually develops under unstable weather conditions, which are characterized by cyclonic activity on this or that territory. Cold weather, usual for winter, is followed from time to time by more warm winter, and sometimes (in the Western regions) by thaws, when snow begins to melt. It means that periodically some territory comes under the influence of cold or warm air masses. But various plots of the area are occupied by these air masses irregularly. When the weather changes from cold to warm, cold air recedes first of all from open plots and stays longer in forests. The approaching warm air, as more light one, does not eject cold air out of the forest but glides above. The weather being less windy, cold air remains in forests for longer period of time, depending on thickness and territory of the forests. This phenomenon is confirmed by low wind speed in forests. On the contrary, when the weather changes from warm to cold, the approaching cold air which is heavier, fills up the forest plots rather quickly. Warm air remains in forests only for a short period of time. Averagely during winter it is more cold in forests than in open areas, which is confirmed by temperature measures. This is shown by Tolsky ([11]), Mütrich ([16]) and others and is confirmed by the investigations of subsequent years, the summary of which is given by Nesterov ([6]) and Kittredge ([14]).

Thus, open areas are subjected to warm air effect more than afforested plots. Therefore, during winter, snow decreasing in open areas is greater than in forests and as a result of this at the end of the winter snowpack accumulation in forests appeare to be larger than in fields, meadows and other open areas, in spite of the fact that some portion of snow stays on the crowns of trees.

At first this phenomenon can be explained by more intense snow melting in open areas during thaws. It is seen in the Western regions of our country. But this explanation cannot be applied to the Eastern regions where thaws are very rare and so insignificant that snow melting dives only small quantity of water. absorbed by snow cover and water return in to the ground is often absent. Here takes place irregularity of snow evaporation in fields and forests.

In forests under the conditions of low air temperatures and wind calm, the evaporation of snow is less than in fields and this causes larger snow accumulation by the end of winter. Even in so damp countries, as Germany is, after Müller-Delisch ([11]) the evaporation of snow in winter in the forests is 35-40 mm less than in open areas.

Generally in winter evaporation measurements of snow surface encounters some methodic resistance, especially durnig snowfalls and snow storms. However, there are some reasons to believe that especially during snow storms, when the smallest snow particles rise into the air, the evaporation may considerably increase.

It is necessary to take into account that in winter snow intensively evaporates from tree crowns, which detain snow during snowfalls.

If the snow is not shaked off the trees by violent winds, the considerable part of it inevitably evaporates and is lost for the snow storage. This can be observed in the cases when during snow stroms and long after them warm weather with weak winds sets in.

It appears that such cases must be taken into consideration while determining the relationship between the difference of snow storage in forests and in fields and the sum of positive air temperatures for the period of snow accumulation.

The sum of positive temperatures should include only those temperatures which were observed in the periods of small amount of snow as well as in the periods of intense snowfall and strong winds which cause shaking the snow of the tree crowns. Fig. 3 shows the relationship between snow storage difference in forests and fields and the sum of the positive air temperatures at 13 houres at the same station, Vassilevichi, Byelorussian SSR.

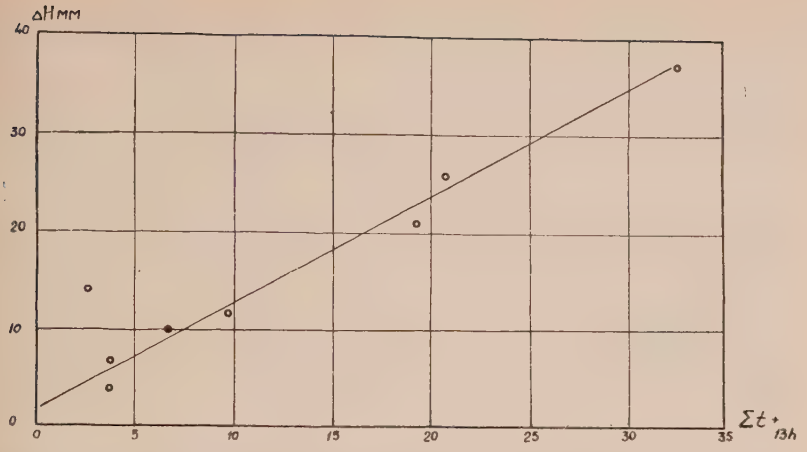

Fig. 3 — The graph of relationship between snow storage difference in forests and in fields (ΔH_{mm}) and the sum of positive air temperatures at 1,00 p.m. for the period of snow accumulation ($\Sigma t^0_{+\,13\,h}$) on the hydrometeorological station of Vassilevichi (Byelorussian SSR).

Calculating the sum of positive temperatures at 13 houres we excluded positive temperatures for those ten snowy days (with precipitation exceeding 10 mm) when mean diurnal wind speed was less than 5 m/sec. On the contrary, positive temperatures for ten days, when it was snowing fast, but when mean diurnal wind speed was more than 5 m/sec, were included in the sum. As a result, there appeared to be definite relationship between snow storage difference in forests and fields and the sum of positive temperatures.

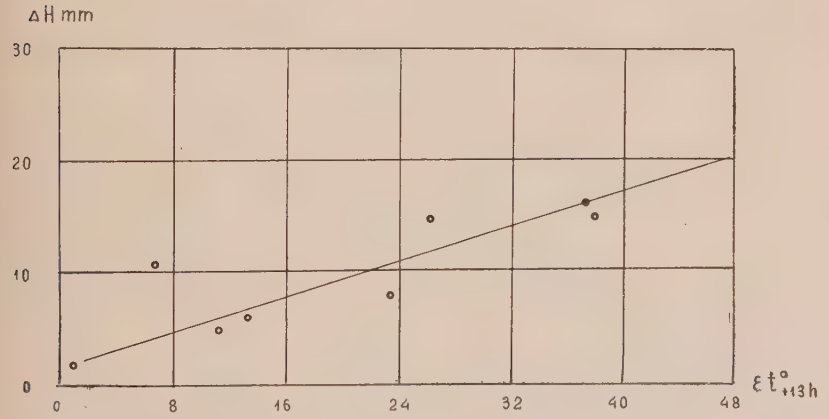

Fig. 4 — The graph of relationship between snow-storage difference in forests and in fields (ΔH mm) and the sum of positive air temperatures at 1,00 p.m. for the period of snow accumulation ($\Sigma t^0_{+\,13\,h}$) on the run-off station of Bolshoe Sareevo (Moscow region).

It may probably be more definite if one consideres not the mean diurnal but maximum wind speed. Besides, in further investigations, the criteria of « much snow» for this or that period are to be defined more exactly.

Analoguously there have been calculated the sum of positive air temperatures for other points of the country. Almost everywhere can be observed more definite relationship between snow-storage difference in forests and fields and the sum of positive temperatures for the period of snow accumulations. Fig. 4 gives such relationship for snow surveying 12 plots (7 field plots and 5 forest plots) of the run-off stations of Bolshoe Sareevo, situated near Moscow. Forest snow-surveying plots are situated in deciduous (birch) and mixed forests consisting of coniferous (fir, pine) and deciduous (birch, asp) species.

These dependencies stress once again the fact that relative growth of snow storage in forests is connected with the rise of temperature in winter. They enable to calculate snow storage in forests on the basis of field snow-surveys and observations of temperature during winter. These relationships once more acertain that the relative increase of snow accumulation in forests is caused by high temperatures during winter period.

Relatively greater snow accumulation in forests is porbably caused by the «horizontal» precipitation as well which, according to some investigations ([2, 11, 13]), may reach considerable size (6-10 per cent and more) especially in mountain areas. But during unstable weather this kind of precipitation also increases. Therefore the relationship between snow accumulation in forests and weather conditions can be somehow explained by «horizontal» precipitation .

Such are the pecularities of forest influence on snow accumulation during winter period when snow storage accumulation prevails over snow melting.

In spring period forests greatly effect the cause of the winter snowpack melding. In the majority of cases snow in firests melts slower than in fields and the melting of snow cover continues for a long time. Snow-storage coefficients greatly increase.

One of the main reasons of this is to be the weakening of direct solar radiation in forests due to overshading, which diminishes the enlightment of snow cover.

For instance, Kusmin's investigations ([4]) have shown that firgroves of mean thickness weaken solar radiation to 2-3 times. The observation of the run-off stations of Bolshoe Sareevo show that thick mixed forests weaken solar radiation in some degree. The effect of solar radiation becomes less in thick forests with young trees and underwood. However, in decidious forests where leaves are shedded in winter it is higher than in coniferous forests but less than in fields.

The weakening of direct solar radiation in forests becomes apparent only in clear sunny weather while in warm cloudy weather its influence naturally is not great. Delay of spring snow melting in forests can be observed in dull weather of cyclonic type as well. So, there is one more reason of delayed snow melting in forests. From our view point it is the same reason, which explains the greater snow acccumulation in forests during winter period, that is, the detain of cold air by forests, especially in the surface layer, where in the presence of young forests and underwood, calm and very low wind speeds prevail. Mass measurements of air temperatures in forests and fields which were conducted in different countries at forest meteorological stations show that in general during snow melting period forest air is 0.5-1.5 and more colder than the air of open plots.

From the view point of forest influence on the river regime during spring floods it is important not only the fact of slower snow melting in forests and the delay as a result of snow fading in woodlands. It is important that this delay can vary in different conditions. It explains different durations of snow melting in forests and fields. For instance, in 1948 at the run-off stations of Bolshoe Sareevo snow in the forests faded 4 days later than in the fields. in 1949-7 days later, in

1947-18 days later and in 1950-11 days later. The same can be observed at other stations. It is clear, that the more snow melting in forests detains from that in fields, the more water-regulation influence of forests can be observed.

The duration of the period between the dates of snow fading does not depend on the total volume of snowpack accumulation but is connected with the snowpack storage correlation in forests and in fields (snow-pack coefficient), which is set in by the end of snow accumulation period.

However the course of weather plays more inportant part in the detain of snow melting in forests in comparison with fields. The earlier ([8]) analysis of cynoptic conditions in various years in the central regions of the European territory of the USSR shows that the largest delay of snow melting and snow fading in forests comparing with that in the fields is observed in the years with unstable weather caused by periodic passing of the cyclones and crest series. Such was the year of 1947, when warm cloudy weather was followed by the fall of temperature, caused by cold air intrusion. In the central regions snow in fields averagely melted away for the period of 11 days and in forests, where snowpack accumulation was 15-20 per cent more and where cold air stayed longer snow melting continued for the period of 29 days.

Slow snow accumulation in forests was favoured by cold weather which lasted for some time after the snow melting. Due to the course of melting, the flood of the Moscow-river expected to be in 1947 very high, was divided into two peaks and ended without great floods. (fig. 5).

This year water-regulation influence of forest showed itself with maximum in the basin of the Moscow-river.

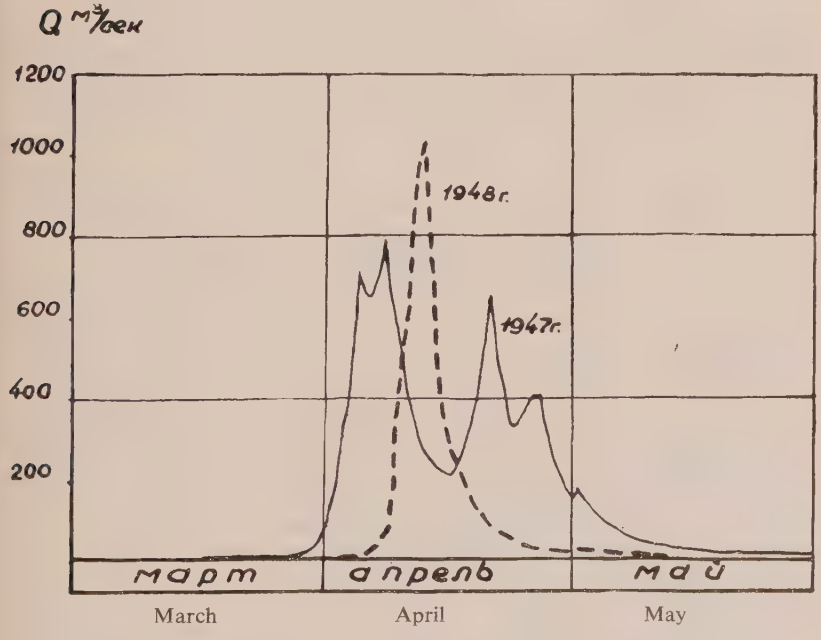

Fig. 5 — The hydrographs of Moscow-river flood by Zvenigorod in 1947, 1948.
Bassin area — 5000 km².
Afforestation — 37 per cent.

To the North and North-East from Moscow, in the same 1947, snow melting began later but, as a result of high temperature, it continued from April 15 without intervals and with the same intensity in forests and in fields. The excess of snowpack accumulation in forests over that in fields averaged not more than 10 per cent. Due to the simultaneous melting, the delay of snow fading even in coniferous (pine-fir) forests was only 2-3 days. It caused very high flood on rivers, including rivers with afforested basins. This is seen on Kostroma-river, the basin of which is afforested on 67 percent. (fig. 6).

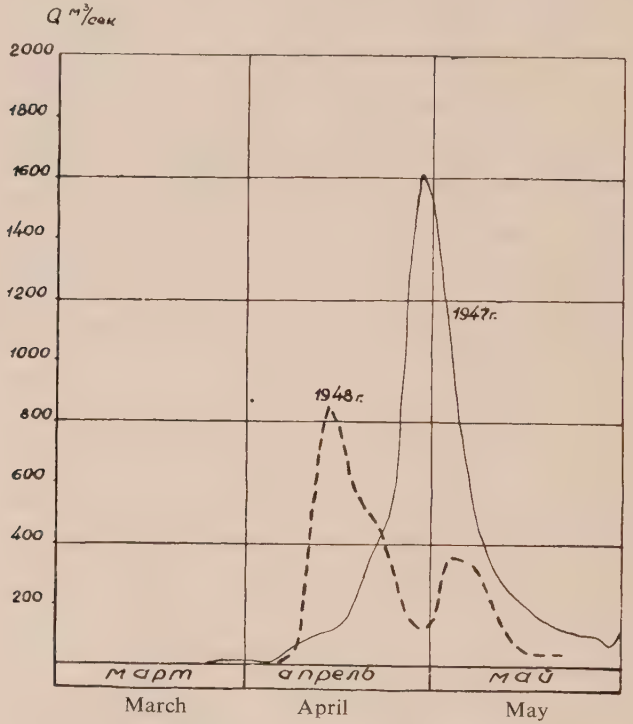

Fig. 6 — The hydrographs of Kostroma-river flood by the town of Bui in 1947, 1948.
Basin area — 8870 km².
Afforestation — 67 per cent.

So in 1947 the forests of the Northern and Eastern regions have not been such waterregulation factors.

Quite another was the cause of snow melting in 1948. By the april of that year a steady anticyclone was formed over many regions of the European territory of the USSR; in the central regions cloudless windless weather set in; the air warmed quickly, in the day time the temperature reached 10° C. The snow melting was caused by warm air, as well as direct solar radiation. Snow melting intensity in deciduous and mixed forests differs little from that in fields. Therefore, snow melted swiftly over the whole area of river basins. As a result, melted snow streamed down simultaneously from all basin plots in forests and in fields. This led to very high floods on some rivers: the Moscow-river, for instance (fig. 5), where it was higher than in 1947, although in 1948 snow storage of the basin was smaller.

As in the Eastern regions, in 1948, the air warmed slower, which was connected with inflow of cold air along the North-Eastern part of the anticyclone, snow melting was not so simultaneous. Snow fading in forests delayed on 8-10 days from snow fading in fields. This caused the reduction of maximums and the formation of the second wave of the flood on rivers (fig. 6).

The mentioned examples clearly show that in some cases forests regulate greatly the run-off in the period of flood, depending on the character of the weather in the period of spring snow melting, in other cases water regulating influence of forests is not so great.

Thus, the well-known statement that forests detain snow cover melting under various weather conditions justifies itself differently. Therefore, in practice, considering the forest cover effect on the river water regime, it is necessary to take into consideration the area and character of the forest in the basin, as well as the general course of weather on the given territory in winter and in spring in the snow-melting period.

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UN DEMI-SIÈCLE D'OBSERVATIONS A L'ÉCHELLE NIVOMÈTRIQUE DE L'EIGER

(*Alpes bernoises*; *station « Eismeer » du Chemin de Fer de la Jungfrau, 3160 m*)

le Dr. PAUL-L. MERCANTON, prof. hon.

Lorsqu'en 1891 F.-A. Forel donna à la glaciologie son célèbre « Rapport sur les variations périodiques des Glaciers des Alpes Suisses » premier terme d'une série annuelle arrivée aujourd'hui et sans lacune à son 77^{me} numéro, le génial précurseur savait pertinemment que le contrôle des avances et des reculs frontaux du glacier ne saurait suffire. Il imposait celui des fluctuations de l'enneigement des hautes régions nourricières de la nappe fluente des glaces. Mais, en ce temps-là encore, l'Alpe demeurait quasi fermée à l'homme; seuls s'y aventuraient quelques intrépides chasseurs, quelques grimpeurs de trempe exceptionnelle et, mais moins haut, bûcherons et forestiers.

La fréquence et les dégats des avalanches renseignaient incidemment sur le degré d'imporatnce de l'enneigement hivernal. Toute visite des hauteurs se faisait à pied en fpulant la neige, avec un équipement insuffisant, des abris rares et précaires et, en cas de malheur, sans secours assurés. Il faut attendre l'aurore du XX^{ème} siècle et l'avènement du ski pour que la montagne s'ouvre à l'alpinisme, avec la multiplication des refuges, l'amélioration de l'équipement individuel et l'enrichissement technique et des moyens de transport. Néammoins Forel s'était bientôt préoccupé de doter la région du névé d'un mode de contrôle de son enneigement. Dans son XI^{ème} Rapport (1890) déjà il proposa l'établissement, en bordure du collecteur glaciaire, aussi en amont que possible de la limite du névé, d'échelles nivométriques ou « nivomètre ». Ce dispositif serait un échelonnement de traits horizontaux, équidistants, peints en couleur vive et indélébile, au flanc d'une paroi aussi verticale que possible et jusqu'à la neige, sur le parcours habituel des alpinistes et bien visible, de loin déjà. Les divisions, distantes d'un demi-mètre par ex., seraient numérotées de bas en haut à partir d'un zéro fictif et censé être déjà enneigé au-dessous du minimum nival; ceci pour retarder le plus longtemps possible la mise hors service du nivomètre en cas de déchaussement par diminution du glacier. Le touriste serait prié de lire le numéro du trait émergeant de la neige et de la consigner avec la date au livre de la cabine la plus proche, ou de le communiquer directement et le plus tôt possible à qui de droit.

Forel dut attendre plus de douze ans la première réalisation de son idée en Suisse. En 1902 seulement, j'eus le privilège d'établir un premier nivomètre, peint au minium, en traits espacés d'un demi-mètre, immédiatement sous le col d'Orny, sur le chemin, alors très fréquenté, amenant au grand plateau glaciaire du Trient et à 3100 m. d'altitude. Cette échelle a fourni durant la belle saison de précieuses données. Les difficultés de sa visite en hiver n'ont guère permis de recueillir des renseignements sur le maximum d'enneigement printanier, jusqu'au moment où d'un avion de chasse passant en rase-mottes devant le rocher, j'ai pu obtenir par la photographie quelques valeurs précieuses de ce maximum. Malheureusement, la déglaciation progressive de nos Alpes a mis le nivomètre d'Orny hors service vers 1932. Il faut bien dire ici que la contribution escomptée des touristes à l'observation de ce nivomètre a été absolument misérable. Heureusement j'ai pu compter sur le zèle des gardiens des cabanes d'Orny et Dupuis.

Un nivomètre établi en 1925 près de la cabane Bertolles (Val d'Hérens), par Mr. A. Renaud n'a, pour la même raison, fait qu'un trop bref service. Il en a été autrement du nivomètre que j'ai peint en 1907 à 3030 m. d'altitude, à la paroi dominant le glacier des Diablerets. Il a été observé année après année par Feu Ernest Reber, au cours d'un millier d'ascensions; ses fils continuent à le surveiller. Malheureusement, sa graduation n'ayant pu être poussée assez haut, l'enneigement annuel n'a guère été convenablement déterminé.

Les conditions d'établissement et d'utilisation du nivomètre sous la station « Eismeer » du Chemin de Fer de la Jungfrau ont été infiniment plus satisfaisants et je lui consacrerai ici la mention qu'il mérite. Préalablement toutefois signalons qu'un ensemble de cinq nivomètres a été établi tout autour de la Concordia pour l'étude hydraulique du bassin de l'Aletsch par le Service fédéral des Eaux (Lutschg) cet ensemble a servi de 1912 à 1923 seulement.

Grâce à l'appui éclairé et dès lors indéfectible de la Cie du Chemin de Fer de la Jungfrau et de son directeur d'alors, le Dr. Liechti, j'ai pu établir, le 3 octobre 1906, à la station « Eismeer », à l'époque tête de ligne, une échelle nivométrique de grand développement et vraiment rationnelle. La station, creusée dans une roche compacte (malm), d'une paroi abrupte (environ 55°) dominant le Grindelwaldervieschерfirn, affluent du Grand Glacier Inférieur du Grindelwald. prend jour vers l'Est par de grandes baies. L'une d'elles dominait une sortie aménagée une quarantaine de mètres au-dessous du niveau du névé. Déhalé sur une sellette, pot de minium et pinceau en mains, j'ai réussi à tracer sur la roche lisse une graduation en degrés d'un demi-mètre et numétoré de bas en haut jusqu'au névé, alors à son minimum et affleurant le degré 3, au-dessus du zéro nivométrique, dont la cote est 3118 m. sur mer.

Constamment surveillé et tenu en bon état, prolongé au besoin par le personnel du Chemin de Fer, avec une ponctualité et un dévouement dignes de tout éloge et que nous en remercions chaleureusement ici, le nivomètre de l'Eismeer a fourni à la glaciologie une série eininterrompue — malgré deuxg uerres — de données uniques en leur genre sur le rythme de l'enneigment et du déneigement d'un collecteur glaciaire dans son encaissement rocheux, à quelque 300 m. en amont de la limite du névé. Force est de ne donner qu'ici un résumé cinquantenaire des milliers de lectures faites. Le voici, limité aux seuls extrêmes de l'enneigement, minimum et maximum en degrés: l'échelle (2 degrés = 1 mètre), minimum moyen 11, compris entre —6 et +26 degrés.

La date moyenne minimum a été le 22 septembre, entre les époques extrêmes : 30 juillet et 28 novembre.

Pour l'enneigement maximum annuel on a : maximum moyen 48 degrés dans les limites 32 et 78 degrés; le maximum est intervenu en moyenne le 25 mars, entre les époques extrêmes 9 janvier et 26 mai.

Le bilan cinquantenaire du nivomètre de l'Eiger — Eismeer se présente ainsi pour la période entière 1906/1956 (50 ans) et en mètres :

somme des enneigements annuels : 996,5 m.
somme des déneigements annuels : 990 m.

soit 6,5 m. au profit de l'accumulation au haut du collecteur de l'affluent W du Glacier Inf. du Grindelwald, dont l'apport au front à travers le chenal Eiger—Mettemberg égale celui de l'affluent Est, venant du Schreckhorn, plus étendu. Nous connaissons des variations de longueur de ce complexe glaciaire durant le même demi-siècle aussi. Il convenait donc de rechercher l'éventuelle corrélation entre la série nivométrique de l'Eismeer et le comportement, contemporain, du front, d'autant plus que la première moitié du demi-siècle en cause a vu le glacier réagir à la courte phase de crue, assez générale chez nous, être 1913-et 1924,.

De 1906 à 1933 le front du Glacier Inférieur a subi des alternances de crue et

de décrue, cette dernière l'emportant toutefois par une crue globale de 675 m. De 1934 à 1956 le retrait a été au contraire continu, atteignant ainsi 4130.

Pour le nivomètre on avait eu les bilans d'enneigement et de déneigement suivants (totaux) :

1906/1933 (27 ans) + 9,5 m.
1934/1956 (23 ans) — 3 m.

Si donc le bilan cinquantenaire est positif encore, par +6,5 m. à la naissance amont du collecteur glaciaire, c'est à la permière période qu'il le doit ; la deuxième ayant été déficitaire en même temps que le retrait du front dissipateur s'affirmait fortement.

Une remarque encore en terminant. L'accumulation hivernale des neiges au pied du nivomètre doit provoquer un accroissement de l'écoulement du névé à cette époque de l'année. J'en ai recherché les indices dans les données du nivomètre sans parvenir à avérer nettement cet effet plausible, lequel ne peut que réduire le chiffre nivométrique. Une telle échelle ne peut donc pas remplacer une balise érigée dans le névé et prolongée au fur et à mesure de son enfouissement.

Le nivomètre n'a pas éveillé assez d'intérêt de la part des alpinistes pour engager à multiplier un dispositif qui n'exigeait pourtant qu'un minimum de frais d'établissement et d'entretien. Notre espoir a été déçu et la nivométrie secondée aujourd'hui par des moyens de transport et de séjour en hautes régions, toujours plus efficaces et abondants, mais surtout organisé pour un travail continu d'équipes de spécialistes rémunérés, est entrée dans une phase nouvelle qui relègue nos vieilles expériences dans l'histoire de l'exploration alpine. Il convenait pourtant de ne pas laisser se perdre les quelques résultats obtenus dans l'enfance de la nivométrie.

On trouvera tous les détails sur les nivomètres cités plus haut et leurs données dans la série des « Rapports Forel », parus annuellement, d'abord dans le « Jahrbuch » du Club Alpin Suisse, puis dans la revue « Les Alpes », leur suite.

ON METAMORPHISM AND HARDENING OF SNOW UNDER CONSTANT PRESSURE AND TEMPERATURE GRADIENT

M. R. DE QUERVAIN
Swiss Snow and Avalanche Research Institute Weissfluhjoch/Davos

ABSTRACT

The different development of a fresh snow layer under a heavy or light load of superimposed snow is well known. It is also known that the change of the structure and in particular the development of strenght is the result of a combined action of pressure, temperature gradient and temperature. The formation of certain types of avalanches is greatly related to this phenomenon, as well as the behaviour of compacted snow banks and underlying strata. In a simple set of experiments carried out in the laboratory, it is snown how certain properties of snow change with time, when kept under various conditions of pressure and temperature.

1. THE PROBLEM

Observations taken outdoors in a great number of snow profiles and over many years have already given important qualitative reference on the general behaviour of snow under various temperature and stress conditions. However, it is rather difficult to draw quantitative conclusions, since natural conditions are variable and mingled with other factors such as wind and radiation. Therefore a set of snow samples was subjected in the laboratory to well defined tempeartures, temperature gradients and loads with the special aim to study the metamorphism of the snow crystals and the hardening of the material as functions of these basic factors.

To illustrate the importance of the problem an example for the natural development of snow shall be presented first. In the winter 1950/51 — the catastrophic avalanche winter of the Alps — a series of snow profiles were taken at Weissfluhjoch at regular intervals (Fig. 1). In this graph snow depth is plotted against time and hardness, snow temperature and snow structure against snow depth. From mid November to mid January the snow depth remained at about 100 cm. Then there was a sudden increase by about 150 cm within several days. During the first period the hardness of the superficial layer, as measured by a cone penetrometer was low and increased only slightly after the fresh snow had imposed an additional load of about 300 kg/m², whereas the new snow itself reached the highest hardness ever measured in dry snow at Weissfluhjoch (250 kg). Already in January the old snow was composed of medium sized, sharp edged grains and represented a high degree of metamorphism. Obviously the fresh fluffy snow was much more susceptible to compaction than the old coarse granular material. Furthermore it seems that temperature and temperature gradient govern the process of solidification together with stress and time. In the early winter the gradients had been considerably higher than after the heavy snow fall. A more impressive story than by hardness figures and crystal sizes is told by the fact that several days after the snow fall had ceased the enormous snow cover was safer than ever before and did not produce any further avalanches. This behaviour must widely be based on the processes which are the subjects of this paper. An other related problem of a certain practical importance is snow compaction on roads and runways.

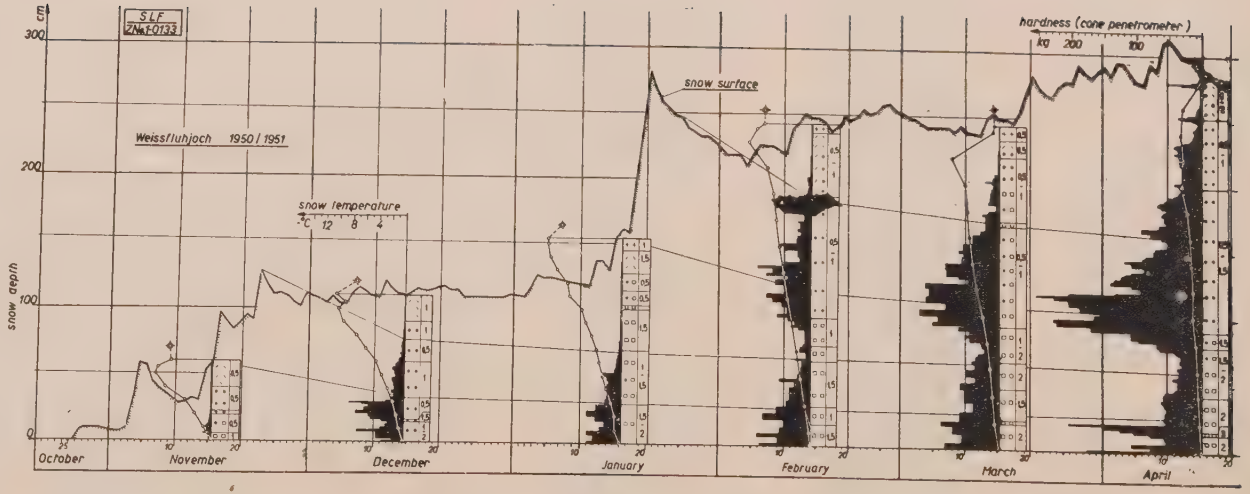


Fig. 1 — Influence of a heavy snow fall on solidification. Development of depth, temperature, hardness and stratification of the snow cover in 1950/51 at Weissfluhjoch (2670 m) in the Swiss Alps. Black areas : hardness profile. Columns : Crystal type and size. (Crosses = *a*, bars = *b*, dots = *c*, squares = *d*, according key to table 1, page 232 ; size in mm).

2. Arrangement of the Experiments

Fresh snow was sieved into 12 wooden boxes (Fig. 2) and stored during 40 days under the following conditions:

Boxes No. 1 *to* 3: temperature constant —5°C, vertical unit loads (without own weight of snow) 4 kg/m², 160 kg/m², 400 kg/m².

Boxes No. 4 *to* 7: temperature constant —13°C, vertical loads same as mentioned above. Box 6 with plastic foils inserted at four levels.

Boxes No. 8 *and* 9: vertical temperature gradient of 0.18°C/cm with bottom at about —4° and surface at about —6°. Vertical loads 4 kg/m², 160 kg/m².

Boxes No. 10 *to* 12: Vertical temperature gradient of 0.70°C/cm with bottom at about —1°C and surface at about —8°C. Vertical loads same as mentioned. Box 12 with plastic foils inserted at four levels.

Details may be taken from Table 1 (left side) and Fig. 2. Fig. 3 shows one box and the electrical equipment.

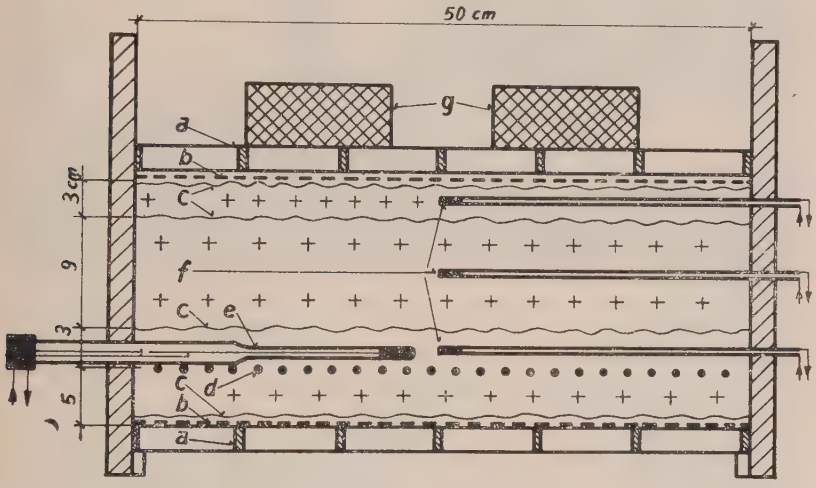


Fig. 2 — Experiment No. 12. Cross section of box.

a = wooden grill
b = wire mesh
c = plastic foil
d = electrical heating grid
e = electrical mercury thermostat
f = thermocouple (Cu — constantan)
g = load
snow layers between foils: crosses.

The temperature gradients in samples 8 to 12 were produced by electrical heating elements (wire grids) and controlled by means of mercury thermostats. As the snow settled during the experiments the gradients did change to a certain extent. This was partly compensated by the increase of the heat conductibility and partly corrected by adjusting the temperatures. A temperature of —5°C was found about half ways between bottom and surface in both sets of gradient experiments. Compared with natural conditions a gradient of 0.18°C/cm is very high, and that of 0.70°C/cm will hardly ever be encountered, except in thin superficial layers being exposed to free radiation.

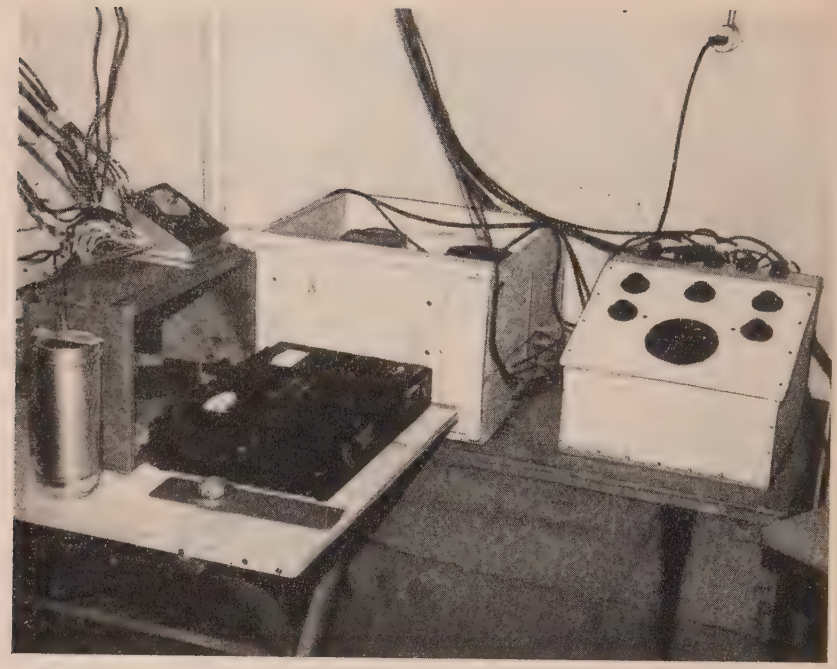


Fig. 3 — Layout of experiments.
Background left: Box No. 12; right: Controll board for temperatures.
Foreground left: Compensation bridge for temperature measurement. (10819).

Thermocouples served for measuring temperatures and temperature gradients.

As the snow was supported and covered by wooden grills and wire mesh, free air circulation was admitted through the snow except for samples 6 and 12. With these two particular experiments it was intended to study the influence of air barriers on the transformation of snow.

All these experiments were of preliminary character. Therefore, to keep them in reasonable proportions, no special stress was laid on extreme constance of temperatures and linearity of temperature gradients. Temperature variations were observed in the range of about $\pm$ 0.5°C and variation of the gradients within 0.15 to 0.25°C/cm for the low gradient and 0.6 to 0.85°C/cm for the high gradient.

3. Methods of Snow Investigations

A specimen of fresh snow and the stored samples were investigated with respect to the following properties:

a) *Settling Process*:
(only stored samples)
The thickness of the layers was measured daily.

b) *Unit Weight*:

Cylinders of 500 cm^3 of snow (58 mm diameter) were taken with their axis horizontally—the samples with a temperature gradient at the —5°C level—and weighed.

c) *Crystal Shape*:

Single crystals picked out from all samples were photographed. To observe the undisturbed structure, orientated thin sections were cut according to the known standard procedure. For a rough specification of the crystal types, the International Snow Classification proved to be sufficient.

d) *Crystal Size*:

The crystal size was determined by two methods: First by measuring the greatest extension of the grains in the photographs and taking the average, and second by a sedimentation procedure from which the grain size distribution by weight is obtained. In this latter case the average grain size can be defined in different ways. We have chosen the value which is exceeded by 50% of the mass of all grains. The two methods do not yield identical results by principle, and, besides this, both are affected in an unfavourable sense by pieces of broken crystals or clustered grains. Thus grain size is a rather inexact figure.

e) *Air Permeability*:

Air permeability is defined as the quantity of air flowing per second through

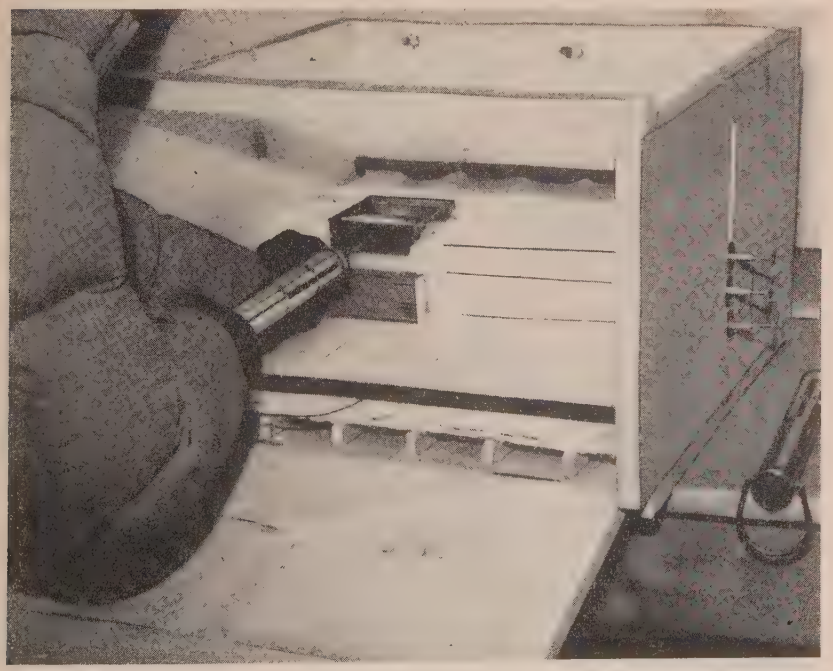


Fig. 4 — Shear test. In opened snow sample heating grid and thermocouples are visible (Experiment No. 11). (10822).

the unit cross section of a snow sample under a unit pressure gradient. It depends on temperature and pressure gradient. Our figures are based on —5°C and a gradient of 0.03 mbar/cm. They were measured for experimental reasons in horizontal direction.

f) *Tensile Strength*:

The tensile strength is measured by rotating a snow specimen with increasing spin until breaking strength is reached by centrifugal force. This was done on the same cylindrical specimens which were used for measuring specific gravity and air permeability. Only two or three parallel tests could be carried out for each material, not a great number considering the high scattering normally observed in this kind of measurements. However, as it was easy to recognize and eliminate any abnormal failures we obtained fair results.

g) *Shear Strength*:

The shear test was taken in horizontal planes by means of a metal frame with 100 cm^2 shear area and a spring balance (Fig.4). The scattering of results is within the range of about $\pm$ 15%.

h) *Hardness*:

For the hardness tests the spring cone penetrometer SLF was used. Although the measured figures—representing the resistance against penetration of a cone tip in kg/cm^2—are only of relative value, this method is useful because it gives consistent results and is free from accidental failures.

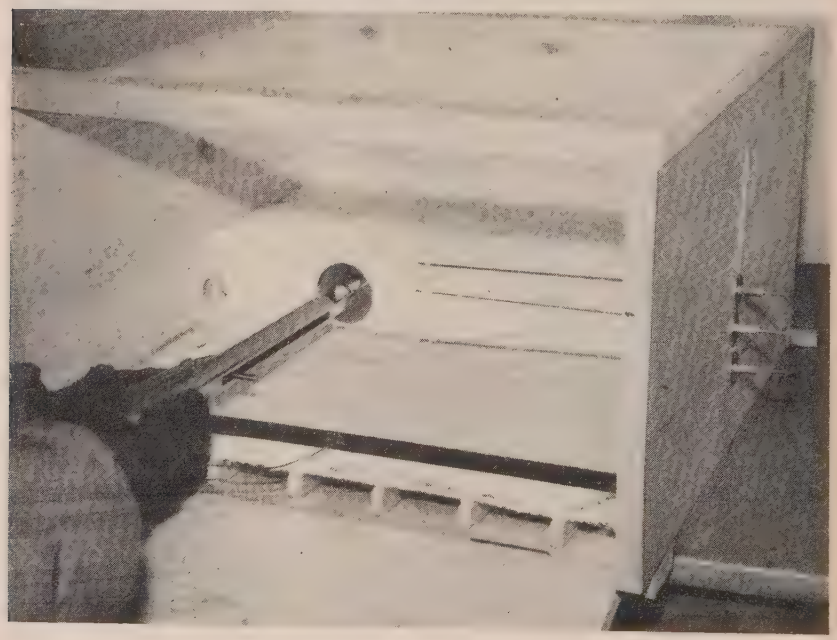


Fig. 5 — Hardness test with spring cone-penetrometer SLF. (10821).

TABLE 1

Properties of snow samples subjected to various temperature and load conditions

No	conditions of experiment				snow properties								
	t	T	grad T	P	H	H′	G	F	D	A	S_T	S_S	R
					Before experiment								
0	0	—4.5	0	0	20	20	190	a,b	1.0/ —	54	7.3	1.6	0.23
					After Experiment								
1	40	— 5	0	4	10.5	14.0	271	b,d	0.70/0.95	113	21.8	4.4	0.44
2				160	9.1	10.7	358	b,c	0.60/0.82	50	95.7	23.0	2.8
3				400	7.3	9.8	388	c	0.49/0.81	40	245.5	25.0	5.9
4	40	—13	0	4	14.0	15.0	254	b,c	0.83/1.11	120	15.5	3.2	0.43
5				160	11.0	11.7	324	b,c	0.59/0.87	72	67.9	8.6	1.9
6				*160	12.0	11.9	320	b,c	0.60/0.87	—	—	9.0	1.5
7				400	9.2	10.5	362	b,c	0.55/0.67	52	85.4	18.4	2.9
8	40	— 5	—0.18	4	12.7	14.8	256	d,e	2.0 /1.7	380	2.4	2.7	0,14
9				160	10.1	11.8	322	d,e	1.5 /1.5	200	16.5	8.6	0.93
10	40	— 5	—0.70	4	14.0	17.2	221	e	5.0 /4.8	410	4.8	1.9	0.28
11				160	12.0	14.3	266	e	5.0 /2.5	270	20.2	12.2	0.92
12				*160	12.9	14.3	265	e	2.8 /1.5	290	20.1	14.5	0.95

* With plastic foils inserted

From all these various tests, whatever their single value is we are able to get a good picture of the changes produced in snow by temperature and stress.

4. Results

A summary of the results is shown in table 1 and Fig. 6 to 20. Most of the figures are averages of a number of measurements. A series of additional observations of secondary interest have been omitted for reasons of shortness.

Key for table 1 :

No	= Number of experiment (No. 0 refers to the fresh snow on March 29th 1957)	
t	= Age of snow at time of investigation	(days)
T	= Mean snow temperature in the middle of sample	(°C)
grad T	= Vertical temperature gradient within snow	(°C/cm)
P	= Vertical unit load during experiment	(kg/m²)
H	= Depth of snow layer at end of experiment (Initial depth of all specimens: 20 cm)	(cm)
H′	= Depth calculated from G, ($H' = H_0 \cdot G_0/G_t$) H′ — H represents roughly the loss by evaporation	(cm)
G	= Unit weight (specific gravity) of snow Average over a layer of 6 cm depth	(kg/m³)
F	= Grain shape: a = dendritic crystals (fluffy snow) b = oblong crystals, some branched c = isometric, rounded crystals d = isometric, edged crystals e = hollow crystals with even facets (depth hoar)	
D	= Grain size. First figure: from photograph Second » : from sedimentation test	(mm)
A	= Air permeability	(cm³/s.cm² per mbar/cm)
S_T	= Tensile strength	(kg/dm²)
S_S	= Shear strength	(kg/dm²)
R	= Hardness measured by spring cone-penetrometer	(kg/cm²)

5. Discussion

a) *Settlement*:

It is interesting to note that the settlement is slowed down by a temperature gradient (compare No. 1 with 8 and 10). Evaporation (i.e. H′ — H) is somewhat more pronounced in presence of a temperature gradient, but not as much as one might expect. In sample No. 6 evaporation has been entirely prevented by the plastic foils, which was not the case in sample 12 with a gradient.

b) *Specific Gravity*:

All samples show a substantial increase in specific gravity after 40 days, the highest value being produced by the heaviest load and the highest temperature (sample No. 3). The smallest increase however is found not in the coldest unloaded specimen No. 4 but in No. 10. This shows again the influence of the temperature gradient. The figures for No. 2 and 7 are very similar, and we conclude that an increase of pressure from 160 kg/m² to 400 kg/m² is fully compensated by dropping the temperature from —5° to —13°. This effect is observed in most of the other properties

so that samples Nr. 2 and 7 consist of nearly identical snow. Specific gravities are also similar at —13° and under the low gradient of 0.18°/cm, But the general appearance of the snow is quite different and so are most of the other properties.

c) *Crystal Shape*:

Using the symbols of the standard snow classification we can characterize the crystals only in a approximate manner. Nevertheless the result is striking, as can be

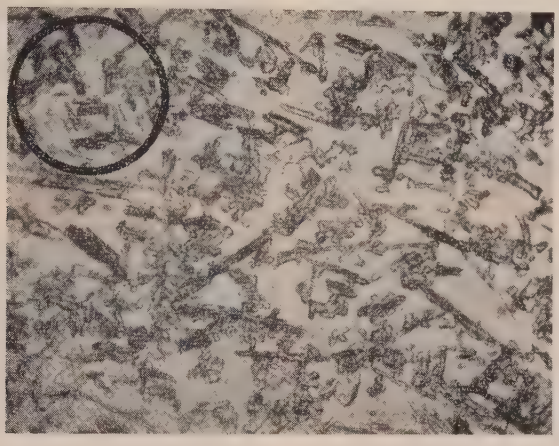


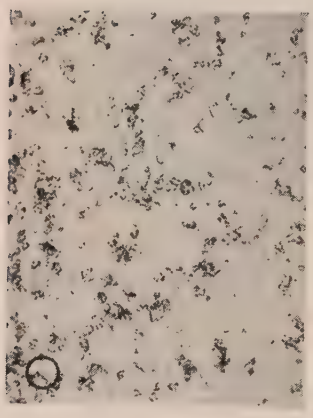
Fig. 6 — Fresh snow before experiment (thin section). (10915). Diameter of circle = 2 mm in nature.

seen from the photographs (Fig. 6 to 20). The upper pictures show single crystals disengaged from the natural structure, the lower ones represent thin sections (0.3 mm thick) of the unchanged structure, seen from the side in the right orintation. In principle two types of crystals are observed. In all those samples which were kept under uniform temperature we find rounded isometric or oblong grains, some still show original branching. At lower temperature we also notice sharp edges and even facets. With increasing stress the single crystals are more and more welded together composing a tight framework.

Under the influences of the lower temperature gradient sharp edged crystals were formed. Some of them having already reached the state of depth hoar with the characteristic cup-like appearance (Fig. 13 and 14). Under the high gradient pure depth hoar was produced (Fig. 15 and 16). The lower the pressure the stronger and more complete was this development. Most depth hoar cups are arranged with their openings towards the bottom as can be recognized from the photographs (13, 14, 15, 16). A single crystal (depth hoar cup) of particular size and beauty was cut through its main axis (Fig. 20). It is subdivided by thin basal lamellas, a typical feature of depth hoar.

In the absence of a temperature gradient, plastic foils did hardly influence the grain shape (Fig. 17). Otherwise they did not prevent the formation of depth hoar in presence of a gradient (Fig. 18). But immediately below the foils dense layers of ice were built up by sublimation (Fig. 19). This may indicate the important vertical vapour transfer under a temperature gradient.

Fig. 9 (3) (10844/10860)
T: —5°C, P: 400 kg/m²



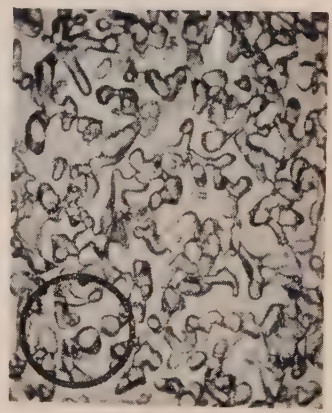
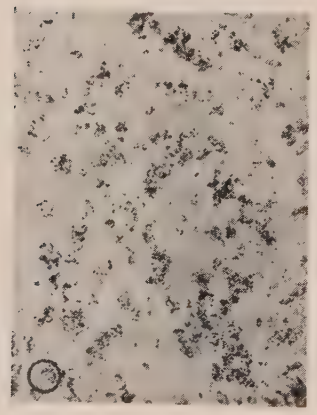


Fig. 8 (2) (10847/10862)
T: —5°C, P: 160 kg/m²



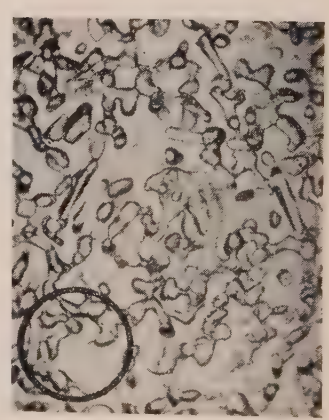
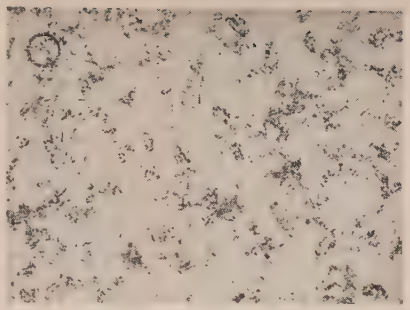


Fig. 7 (1) (10843/10912)
T: —5°C, P: 4 kg/m²

Fig. 7 to 18 — Snow crystals taken from samples after end of experiments. Upper picture: Single crystals loosened from natural structure. Lower picture: Thin section through natural structure (0,3 mm thick). Side view, right orientation (bottom down).
Figure in circle refers to No. of experiment. T = Temperature. P = Unit load. Diameter of circle on photograph corresponds to 2 mm in nature.



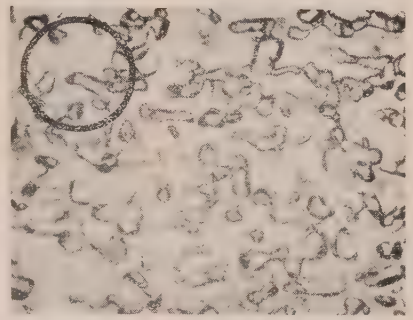
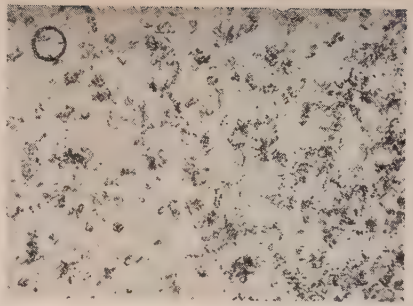


Fig. 10 (4) (10848/10910)
T: —13°C , P: 4 kg/m²



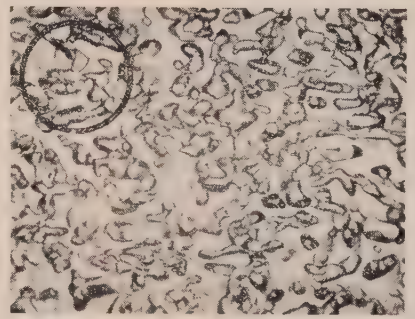
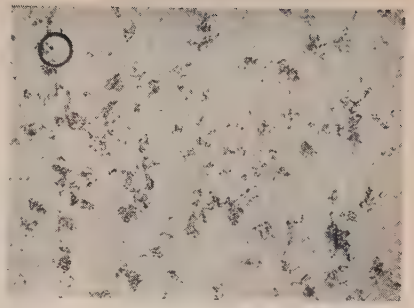


Fig. 11 (5) (10852/10867)
T: —13°C, P: 160 kg/m²



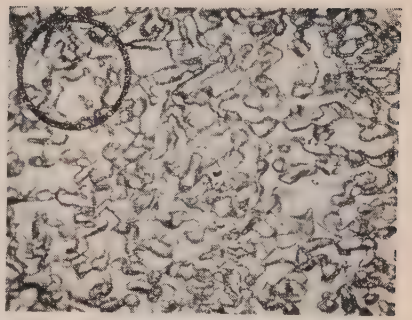


Fig. 12 (7) (10795/10907)
T: —13°C, P: 400 kg/m²

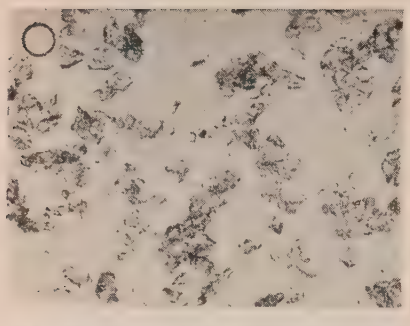
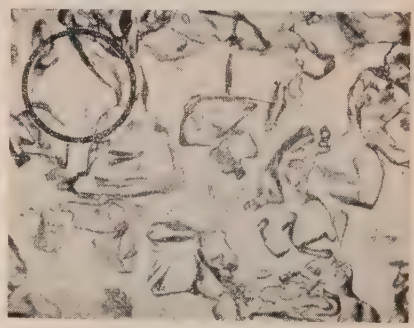


Fig. 13 (8) (10809/10896)
grad T: 0.18°C/cm, P: 4 kg/m²

Fig. 14 (9) (10803/10865)
grad T: 0.18°C/cm, P: 160 kg/m²



d) *Crystal Size*:

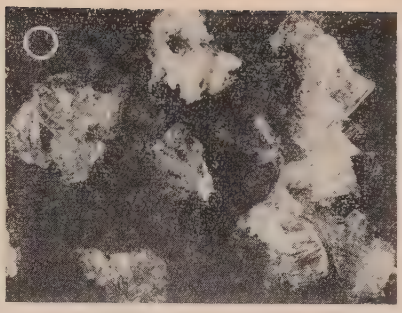
For fine granular snow we get a slightly smaller mean crystal size by the photographic method than by sedimentation. With depth hoar it is just the reverse, because a certain selection is introduced by eliminating splinters of evidently broken crystals.

From the fresh snow of not well defined crystal size first a fine granular material with a crystal size of less than 1 mm is obtained. In the absence of a temperature gradient only a minor growth will take place. This is evidently hampered by compression. A quite different order of magnitude appears in the samples with a gradient. The number of individual crystals drops to a small fraction and the size of the crystals lies between 1.5 and 5 mm. Contrary to our expectations we notice hardly any influence of the load. Perhaps, however, the experiments lasted too long so that something like a final state of growth was reached in all the samples with gradient.

Plastic foils did not affect crystal growth in the constant temperature samples whereas in the specimens with gradient the growth was reduced.

e) *Air Permeability*:

Except in three loaded isothermic samples the initial permeability increased. Again the highest rise was measured in the samples with gradient — a result consistent with the development of specific gravity and crystal size. It must be said that the high figures above 200 were probably outside the normal range of the measuring instrument. With proper adjustment they would have become even higher.



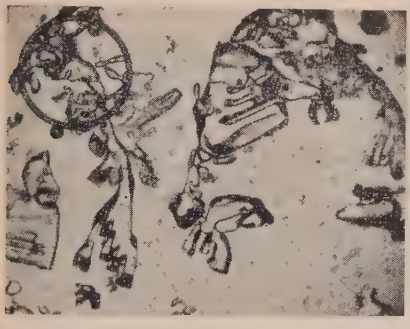


Fig. 15 (10) (10838/10901)
grad T: 0.70°C/cm, P: 4 kg/m²

Fig. 16 (11) (10823/10898)
grad T: 0.70°C/cm, P: 160 kg/m²

Between the foils a surprisingly high air permeability was observed (No. 12), probably due to the fact that we have measured the air flow between two foils perpendicularly to the direction of the gradient. In any case with a temperature gradient air permeability is raised and this in return promotes metamorphism and loosening of the structure.

f) *Tensile Strength*:

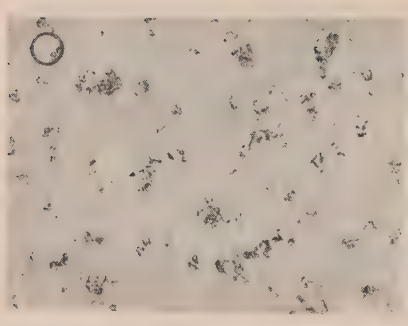
Starting with a figure of 7.3 kg/dm² — wich is rather high for fresh snow and certainly influenced by the sieving operation—we notice a general trend towards solidification. In the isothermic experiments a pronounced rise in strength is produced which becomes greater with increasing load. Colder samples lag behind the warmer ones in strength. Again a parallel behaviour between No. 2 and 7 is observed.

The top value of 145.5 kg/dm² (No. 3) is high, but has been surpassed now and then under natural conditions, for example in the hardest layers of the profile shown in fig. 1 (165 kg/dm²).

With a temperature gradient only the unloaded samples show a loss of tensile strength. The others, although composed of depth hoar, are remarkably strong.

g) *Shear Strength*:

On the whole shear strength follows the same trend as tensile strength. Due to a considerable scattering of the measurements and probably to the anisotropic



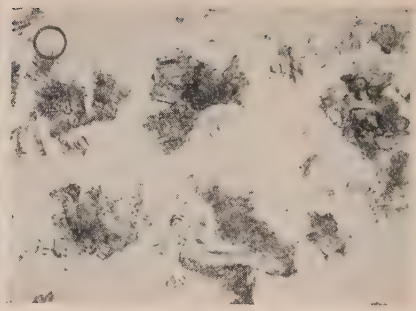
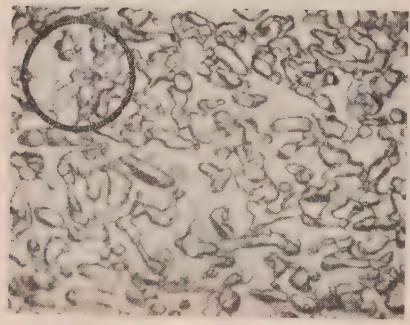


Fig. 17 (6) (10856/10864)
T: —13°C, P: 160 kg/m²



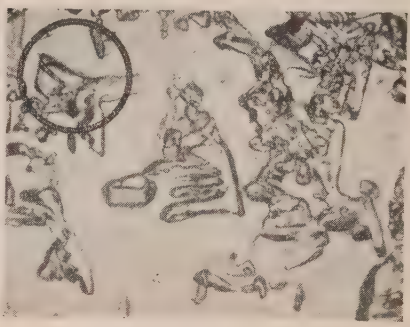


Fig. 18 (12) (10828/10884)
grad T: 0.7°C/cm, P: 160 kg/m²

Fig. 19 — Experiment No. 12. Dense snow layer formed by sublimation underneath third plastic foil (from bottom), seen from the side (thin section). (10903).

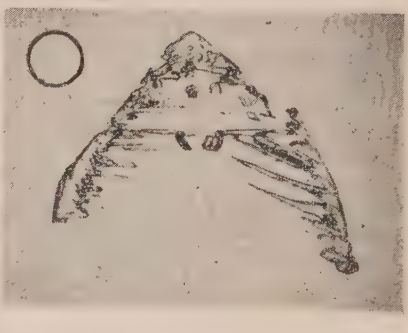


Fig. 20 — Single depth hoar crystal from experiment No. 10. Thin section through main axis of cup (slightly damaged). (10920).

texture of the snow we cannot, however, recognize a sharp correlation between the two propreties.

h) *Hardness*:

Hardness also goes parallel with tensile strength but differently in samples No. 2 to 7 and 8 to 12. In all fine granular specimens, except No. 1 the ratio S_T/R (in the given units) is found between 25 and 36, whereas for the depth hoar samples figures from 17 to 22 are computed. This points to a higher rigidity of the depth hoar structure and perhaps to dynamic effects being involved in our particular hardness test (mass dislocation).

5. Summary and general conclusions

Snow samples were stored for 40 days under various steady conditions of stress, temperature and temperature gradient. A set of measurements taken before and after the experiment verified the general behaviour observed in nature and clarified the separate and combined action of temperature and pressure on snow. In general terms the conclusions are as follows.

a) *In the absence of a temperature gradient* fresh snow is transformed into a fine granular material. Hardness and strength as well as specific gravity increase considerably with a rise of temperature and load.

b) *In the presence of a temperature gradient* fresh snow is transfermed into a coarse granular material — under the condition of the experiments mainly to depth hoar. The development of hardness and strength is checked by a temperature gradient. An increase of strength does not occur unless a certain substantial load is applied. Otherwise even a drop of strength may result. The higher the temperature gradient, the sooner the transformation is achieved.

c) *Specific gravity, strength, and crystalline features* are interrelated to a certain extent. There are, however, snow samples which have similar specific gravity, yet quite different strength, and samples of different crystalline appearance may agree in more than one mechanical property.

d) From the experiments with the *plastic foils* inserted in the snow we conclude that a considerable vapour transfer is produced by the temperature gradient. As for the formation of depth hoar a certain local turnover of incoming and outgoing vapour is necessary. But in a strong gradient depth hoar can develop between foils (or ice sheets in nature) without convection of large scale.

DIE BERECHNUNG DER EISSTANDTERMINE UND TERMINE DER REINIGUNG DER STAUSEEN VOM EIS

W. W. PIOTROWITSCH

SUMMARY

Water reservoirs built on rivers usually freeze and are cleared of ice in other periods than these rivers before the construction of water reservoirs. Freezing and clearing periods of the water reservoirs are established by calculations of a heat balance determined with allowance for the depth and size of the reservoirs. The author has used an equation which unites a heat reserve in the water reservoir and a heat exchange with surrounding atmosphere.

The equation was solved by a «day-by-day» system which permitted to take into account variable meteorological factors.

A comparison was made between observation data on ice drift and complete freezing on the Tsymlyansk, Rybinsk and Dnieper water reservoirs as well as on the Pskov Lake for 66 determinations in different years.

It was found that on open parts of large water reservoirs with average depths less than 5—6 m, time of lowering water temperature to 0° and periods of complete freezing are within the error ± 1—2 days. At narrow and comparatively deep water reservoirs (the Klyazma water reservoirs on the Moscow-Volga Canal, Dnieper Reservoir, Pereborsky Gulf of the Rybinsk Reservoir) with average depths from 9 to 16 m, complete freezing precedes the cooling of water to 0° by a few days. A dependance of the sum of average daily air temperatures below zero was found for the day of the freesing on average depths of small areas of lakes. This dependence permits to compute average dates of complete freezing on future water reservoirs from evidence of perennial observations on the air temperature in the same region.

Annual periods of ice clearing of water reservoirs were computed by a dependence of a cold reserve in the air before thawing by a heat exchange of the ice-and-snow surface of the water-reservoir in spring and by standard meteorological observations.

Observations have shown that a large amount of solar radiation penetrates into the spring ice and the bottom surface of the ice cover noticeably thaws, especially during ice drifting.

The methods of computing the clearing periods were checked up on the Rybinsk, Tsymlyansk, Dnieper and other water reservoirs as well as on the Pskov, Chudskoye and Ilmen Lakes for 43 years. Thus, it was shown that the computation periods coincide with a period of a mass ice clearing of large water reservoirs. Their final clearing is 2—3 days later than the computed date.

From evidence of perennial observations on the clearing periods of the lakes in the USSR, there has been established a graphical dependence between the sum of positive average daily temperatures of the air on the day of clearing and that of average monthly temperatures of the air in winter time. This dependence makes it possible to compute average dates of clearing the water reservoirs according to evidence of perennial observations of the air temperature in the same region.

Für die Ermittlung der Eisstand und Reinigungstermine der Wasserbehälten und zwar der Seen und Stauseen und bei der Ausarbeitung der Methodik für die Prognosierung dieser Termine nutzt man heutzutage vieljährige Eisbeobachtungen der Meßstellen aus. Aber für viele Seen und Stauseen fehlen solche Angaben. Daraus entstehen grosse Schwierigkeiten bie der Zusammenstellung der Prognosen und anderer Auskunften über das Eisregime für solche Seen und Stauseen. Besonders wichtig ist diese Frage heute, wann neue grossen Stauseen an der Wolga, Kama und a. in Gang gebracht werden. Die Mehrheit dieser Stauseen schafft man dort, wo es keine Wasserbehälter gebe, die im Verhalten zum Eisregime Analoge den errichteten sein würden.

Im Zusammenhang damit werden seit dem Jahre 1952 im Zentralinstitut der Prognosen Forschungsarbeiten mit Absicht der Methodikausarbeitung der langfristigen Vorhersage der Eisstand und Reinigungstermine der neugeschaffenen Stauseen durchgeführt. Die erste Etappe der Arbeit wurde der Ausarbeitung einer Rechnungsmethodik dieser Termine gewidmet. Diese Berechnungen, wenn sie gelungen würden, hätten eine Vorstellung von der Regimisete (Mittel und Jahrestermine) bei der bestimmten meteorologischen Verhältnissen einer ganzen Reihe der vergangenen Jahre geben können.

Eine kurze Wiedergabe der Forschungsergebnisse in dieser Richtung wird in diesem Artikel gegeben.

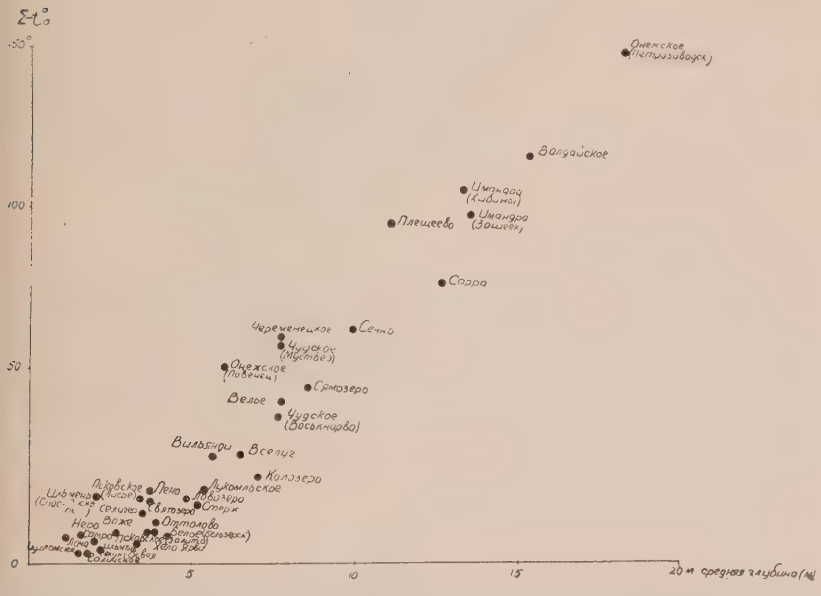
Die Methodikausarbeitung der langfristigen Vorhersage der Eisstand und Reinigungstermine ist jetzt im Stadium der Forschung. Diese Forschungen werden auf dem Basis einer ganzen Reihe der berechneten vieljährigen Termine dieser Erscheinungen durchgeführt.

Die Berechnung der Eisstand- und Reinigungsmitteltermine der Stausen.

Den Berechnungen der mittleren Eisstand und Reinigungstermine der Stauseen wurde die vereinfachte Methode der Wärmebilanz zugrunde gelegt. Für die Ermittlung der Mitteltermine des Eisstandes ist folgende Abhängigkeit festgestellt:

$$\Sigma - t_a = f(H), \qquad (1)$$

wo $\Sigma - t_a$ ist eine Summe negativer Mitteltemperaturen der Luft während vierundzwanzig Stunden aufs Eisstandsdatum, die für die Periode nach dem Herbstübergang der Lufttemperatur über 0° berechnet wurde.



Zeichn. 1 — Der Zusammenhang zwischen der Summe negativer Tagesmitteltemperaturen der Luft für den Eisstanddatum und der mittleren Tiefen der Seen. Die Namen der Seen sind neben den Punkten gegeben.

Diese Summe charakterisiert die Wärmeabgabe eines Wasserbehälters. H ist eine Mitteltiefe des Sees oder Stausees im beobachteten Gebiet. Das ist eine Kennziffer der Wärmemenge eines Wasserbehälters im Falle eines Temperaturrückganges.

Auf der Zeichnung 1 ist die ermittelte Abhängigkeit für die Seen von UdSSR gezeigt. Für die Berechnung dieser Abhängigkeit wurden die Angaben für solcher Seen ausgenutzt, für welche eine grosse Reihe (nicht weniger als 15 Jahre) von Eis und Tiefenbeobachtungen der Meßstellen gesammelt wurde. Für diese Seen wurden Mitteltermine des Eisstandes durch die Bestimmung eines arithmetischen Mittelwertes der jährlichen Angaben bestimmt.

Leider sind die Tiefenangaben nicht für alle Seen bekannt. Für viele Seen, auf welchen mehrere Jahre lang die Eisbeoabchtungen durchgeführt wurden, fehlen solche Tiefenangaben. Infolgedessen hat die Abhängigkeit nicht so viel Punkte.

Die Zeichnung 1 zeigt, daß es eine gute Verbindung zwischen der Summe negatver Lufttemperaturen zum Eistandtermin und der Seentiefe der Nordwest und Zentralgebieten des Europäischen Teils der UdSSR, Karelien und Kola Halbinsel existiere.

Diese Abhängigkeit gibt die Möglichkeit die Eisstandmitteltermine zu ermitteln. Dazu bestimmt man mit Hilfe von Mitteltiefenangaben eines Stausees und der Abhängigkeit (1) die für die Bestimmung des Eisstandes notwendige Summe negativer Lufttemperaturen. Der vieljährige Lufttemperaturengang während vierundzwanzig Stunden wird graphisch dargestellt. Für diese Darstellung nutzt man die Angaben einer nächstliegenden meteorologischen Station aus. Nach der graphischen Darstellung wird das Datum, auf welches sich die notwendige Summe negativer Lufttemperaturen sammelt bestimmt. Dieses Datum gibt den Eisstandmitteltermin des Wasserbehälters.

Man muss unterstreichen, dass die Angaben der auf den Inseln oder auf den Ufern grösser Seen liegenden meteorologischen Stationen bei der Zusammenstellung der Abhängigkeit (1) nicht ausgenutzt wurden. Darum kann man die Angaben der Beobachtungen dieser Stationen für die Bestimmung der Mitteltermine nicht anwenden.

Die Zerstreutheit der Punkte auf der ermittelten Abhängigkeit ergibt Maximum $\pm$ 3 Tage. Diese Zerstreutheit ist hauptsächlich durch nicht ganz genaue Tiefenangaben hervorgerufen. Sehr wichtig für das Eisregime sind auch Klimabesonderheiten der Seen, die nur durch negative Lufttemperaturen über dem Boden nicht zu charakterisieren sind. Eine grosse Bedeutung für die Klimaformierung über einem Wasserbehälter hat die Seenbreite die einen grossen Einfluss auf die Lage der Verbindungspunkte ausübt.

Man muss anmerken, dass die Abhängigkeit (1) für solche Seen festgestellt ist, deren Tiefe nicht grösser als 18 m sei.

In diesem Tiefendiapason ist die Verbindung zur gradlinigen sehr nahe. Mit der Tiefenzunahme wird die Gradlinigkeit zerstört. Z.B. : die Angaben für den Zentralteil von Onegha See, wo die Tiefe 50 m beträgt, liegen weit rechts von der Gesamtverbindungslinie. Das weist darauf hin, dass die Wärmemenge, die im Herbst nicht tiefe Seen verlieren, der Tiefe proportional sind, und die Mischung die ganze Tiefe eines Sees umfasst. In einem tiefen Wasserbehälter verschwindet diese gerade Proportionalität. Erstens ist es darum, dass die Mischung zum Eisstandtermin nur obere Schichten umfasst.

Für die Berechnung der Mitteltermine der Reinigung der Stauseen vom Eis wurde folgende Abhängigkeit festgestellt :

$$\Sigma + t_a = f(\Sigma - \theta) \qquad (2)$$

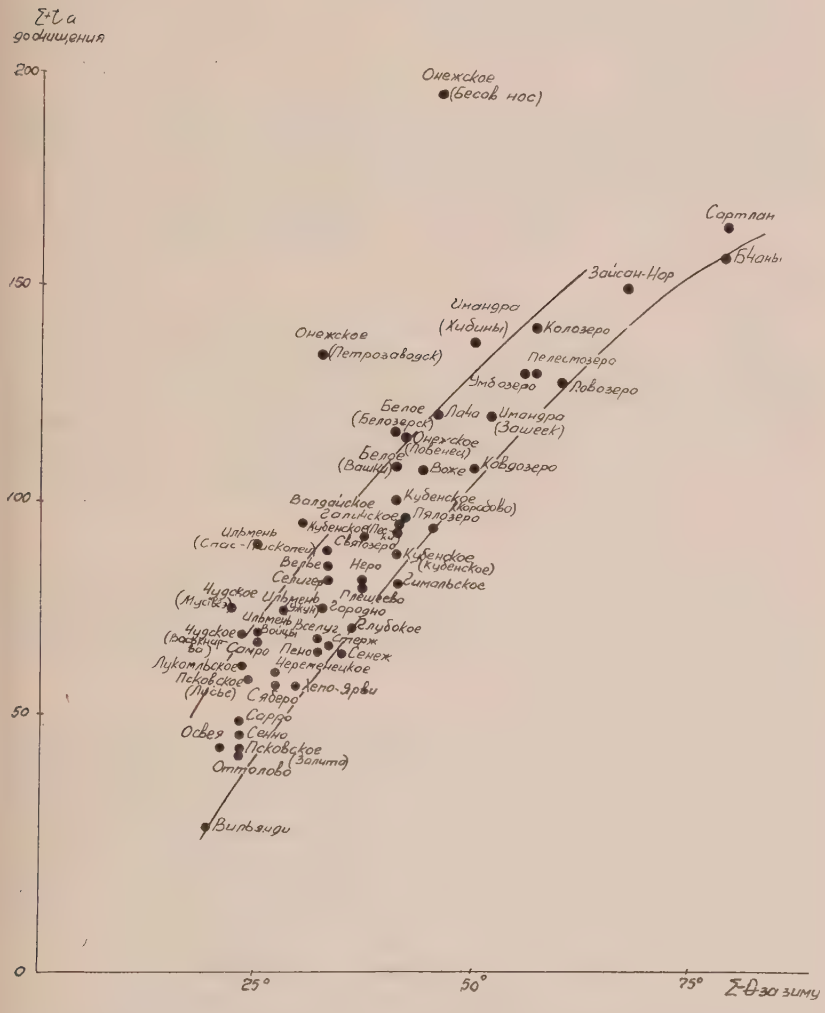
wo $\Sigma + t_a$ ist eine summe positiver Mitteltemperaturen der Luft während vierundzwanzig Stunden zum Reinigungstermin eines Stausees, die seit dem Temperaturübergang, über 0° berechnet wurde. Das ist ein Merkmal der Wärmezufur zum Eis.

$\Sigma - \theta$ ist eine Summe negativer mittelmonatlichen Lufttemperaturen für die

Eisstandperiode. Diese Summe charakterisiert maximale Eisdicke vor dem Tauenanfang.

Für die Feststellung der Art dieser Abhängigkeit wurden die Angaben für solche Seen der UdSSR ausgenutzt, die eine Reihe von Meßstellenbeobachtungen mehr als 15 Jahre hätten. *Die Reinigungstermine* eines Wasserbehälters *vom Eis* wurden durch die Bestimmung eines arithmetischen Mittelwertes der jährlichen Reinigunstermine ermittelt,

Die Entfernung der Punkte (Zeichn. 2) von der Zentralverbindunglinie entspricht



Zeichn. 2 — Der Zusammenhang zwischen der Summe positiver Tagesmitteltemperaturen der Luft für den Reinigungsdatum der Seen vom Eis und der Summe negativer mittelmonatlichen Lufttemperaturen für den Winter. Die Namen der Seen sind neben den Punkten gegeben.

± 2 Tagen. Die Berechnung der Reinigungsmitteltermine eines Stausees vom Eis wird folgenderweise durchgeführt. Man summiert negative mittelmonatlichen Lufttemperaturen für die Eisstandperiode nach den Angaben einer nächsten zu Lande liegenden repräsentativen meteorlogischen Station.

Mit Hilfe dieser Grösse findet man auf der Zeichnung 2 die für die Reinigung vom Eis notwendige Summe der Lufgtmitteltemperaturen während vierundzwanzig Stunden. Auf der graphischen Darstellung des vieljährigen Ganges der Lufttemperaturen in vierundzwanzig Stunden nach dem Nullübergang in Frühling findet man das Datum, zu welchen sich die für die Reinigung vom Eis notwendige positive Temperatursumme sammelt. Dieses Datum wird dem mittleren *Reinigungsdatum* eines Stausees *vom Eis* entsprechen.

Die Art der Abhängigkeit (2) erinnert an eine Kurve des Eisanwachsens im Zusammenhang mit einer sich vergrössernden Summe negativer Lufttemperaturen im Laufe des Winters.

Die Ausnutzung der beobachteten maximalen Eisdicken anstatt $\Sigma - \theta$ gab die Abhängigkeit mit einer viel grösseren Zerstreutheit der Punkte als die ermittelte. Das ist dadurch zu erklären, dass die maximalen Eisdicken auf Grund einer kurzen Beobachtungsreihe bestimmt wurden und darum nicht repräsentativ sind.

Die Zerstreutheit der Punkte auf der ermittelten Abhängigkeit ist durch eine Reihe von Ursachen hervorgerufen. Auf die Reinigungstermine übt einen Einfluss die Breite des Wasserbehälters aus, die eine grosse Bedeutung in der Klimaformierung über den Seen hat.

Die Seen mit einer grossen Oberfläche (Tschudskoje, Ladoga-See und der Zentralteil von Onega-See) reinigen sich vom Eis viel später als kleinere Seen in diesem *Gebiet*. Die Ursache dieser Erscheinung ist von zweierlei Art. Es besteht die Möglichkeit einer Anhäufung von Eismassen auf den grossen Seen im Winter. Das erfolgt nach dem Eisdeckeneinbruch und bei der Packeisbildung. Andererseits kühlt der mit Eis bedeckte See über ihm liegende Luftmasse viel stärker, darum erfolgt hier das Eistauen zu langsam. Daraus folgt, dass die Abhängigkeit für grosse Stauseen, z.B. für den Zentralteil von Rybinskoje Stause, frühere Reinigungstermine in Vergleich zu den beobachteten zeigt.

Auf die Reinigungstermine übt gewissermassen Durchflussgrad einen Einfluss aus. Der Versuch mit Absicht einer Wertung dieses Faktors wurde von uns nach den Angaben über dem Durchflussgrad und über die Termine des Eisbruches für die Seen Ilmen, Tschudskoje, Pskowskoje, Sseliger, Imandra u.a. gemacht. Es stellte sich heraus, dass der Einfluss der Durchflussverschiedenheit dieser Seen auf die Reinigungstermine gering ist.

Um eine genauere Vorstellung über die Eisstand und Reinigungsmitteltermine der Stauseen vom Eis zu bekommen, muss man die volle Wärmebilanz der Wasserabkühlung im Herbst, des Eisanwachsens und seines Tauens mit der Berechnung des Einflusses des Wasserbehälters auf die meteorologischen Verhältnisse berücksichtigen. Aber der Wert solcher Berechnungen ist nicht gross, wiel die Bestimmungsgenauigkeit der Data von betrachteten Eiserscheinungen wegen einer begrenzten Periode der Beobachtungen (15-25 Jahre) für die Mehrheit der Wasserbehälter beschränkt ist. Die ermittelten Mitteldata entsprechen der Beobachtungenperiode auf den Seen in den Jahren 1930-1953.

Die Bestimmung jährlicher Eisstandtermine.

Die jährlichen Eisstandtermine auf den neugeschaffenen oder zukunftigen Stauseen in den hydrometeorologischen Verhältnissen voriger Jahre wurden nach

der Wärmebilanzmethode mit der Berechnung von Besonderheiten des Stausees bestimmt.

Zuerst wurde der herbstliche Temperaturgang des Wassers (T) eines Wasserbehälters nach der Formel

$$dT/d\tau = \sigma/H \tag{3}$$

ermittelt. (1)

Die gleichartige Formel wurde von O. Devik (1) für die Berechnung der Wasserabkühlung des Flusses Glomm (Norwegen) vorgeschlagen.

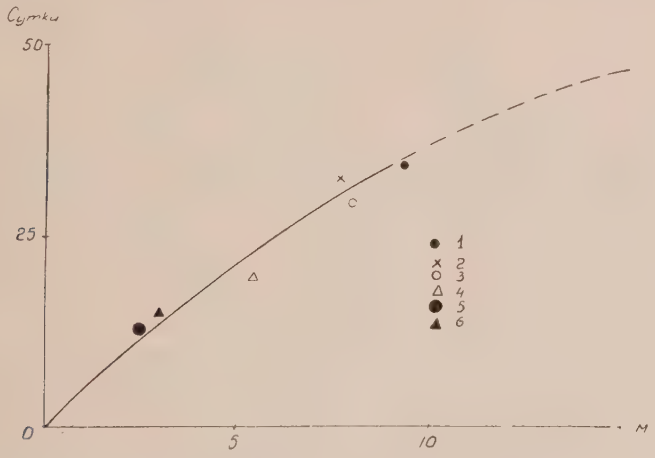
Diese Formel stellt eine differentiale Gleichung dar, da der Wärmeumsatz mit dem umliegenden Raum (σ) und in einigen Fällen auch die Mitteltiefe des Wasserbehälters (H) sich im Laufe der Zeit verändern. Die Lösung der Gleichung (3) in allgemeiner Art ist darum nicht möglich, dass der Gang des Wärmeumzatzes in der Zeit (τ) analytisch nicht auszudrücken ist.

Höchst verschiedene und stark veränderliche Faktoren des Wärmeumsatzes kann man nur bei der Voraussetzung der Unveränderlichkeit der Wassertemperaturen berücksichtigen. Ebenso wie die Wassertemperatur, muss während einer bestimmten Periode auch die Tiefe des Wasserbehälters unveränderlich bleiben. Als solche Zeitabschnitte haben wir Tag und Nacht angenommen. Bei diesen Bedingungen gibt die Lösung der Gleichung (3):

$$T_2 = T_1 + \frac{\sigma\tau}{H} \tag{4}$$

wo T_1 und T_2 ist die Anfangs und Endtemperatur des Wassers auf dem gegebenen Gebiet mit der Mitteltiefe H ;

σ ist summarer Wärmeumsatz des Wassers mit dem umliegenden Raum (Kal/cm²/24 Stunden), mit Hilfe konstanter Wassertemperatur (T_n) bestimmt. Für die nächsten 24 Stunden wurde die Wassertemperatur dem T_2 gleich und unveranderlich ange-



Zeichn. 3 — Die Periode für Bildung des Mittelwertes der Lufttemperaturen, welche für die Bestimmung der Wassertemperatur eines Stausees nötig ist.
1. Die Pirogowski Flussstrecke des Kljasminskoje Stausees.
2. Nebendammgebiet des Iwanjkowskoje Stausees.
3. Pereborski Bucht des Rybinskoje Stausees.
4. Der Insel Mologa des Rybinskoje Stausees.
5. Drei Seen in Oserki bei Leningrad.
6. Die Flüsse Wolga und Kama.

(1) Dichte und Wärme Kapazität des Wassers sind den 1 gleich angenommen.

nommen. Mittels dieser Temperatur wurde der Wärmeumsatz während vierundzwanzig Stunden berechnet; dann wurde wieder die Wassertemperatur für die nächsten vierundzwanzig Stunden berechnet, u.s.w. von Tag zu Tag bis zum Temperaturübergang über 0°. Auf solche Weise ergibt sich der Wassertemperaturgang in der Zeit.

Empirisch wurden die Fehlergrössen, die bei der Berechnung der Wassertemperatur als Resultat der Voraussetzung der Wassertemperaturunveränderlichkeit entstehen, festgestellt. Es stellte sich heraus, dass auf die Fehlergrösse hauptsächlich die Tiefe des Wasserbehälters einen Einfluss ausübt. Bei den Tiefengrössen mehr als 5 m (die für die Stauseen gewönlich sind) kann man die Berechnungsfehler vernachlässigen.

Das ist einfach zu erklären : bei einer kleinen Tiefe (weniger als 5 m) verändert sich die Wassertemperatur sehr schnell, und nähert sich der Lufttemperatur, der Wärmeumsatz aber verringert sich.

Wenn die Tiefen eines Wasserbehälters gross sind, verändert sich die Wassertemperatur während vierundzwanzig Stunden sehr wenig, und die Berechnung des Wärmeumzatzes nach einer konstanten Wassertemperatur recht genau wird.

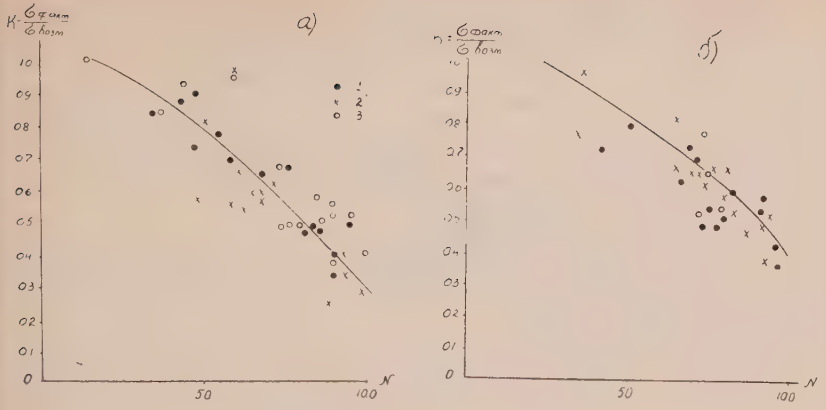
Für die Berechnung nach der Formel (4) ist es notwendig, die Anfangstemperatur des Wassers (T_1) zu haben. Wir ermitteln diese Temperatur durch die Bestimmung der Luftmitteltemperatur für einen vorhergehenden Zeitabschnitt. Die für die Ermittlung notwendige Periode wurde mittels der Gegenüberstellung der beobachteten Mitteltemperaturen des Wassers einiger Stauseen der mittlereren Lufttemperaturen für verschieden Zeitabschnitte entschieden.

Die Periode, für welche man die Mittelwerte bestimmt, hängt von der Mtiteltiefe des Wasserbehälters ab (Zeichn. 3) : je tiefer das Wasserbecken sei desto für eine längere Zeitperiode man die Mittelwerte der Lufttemperatur bestimmen muss; dementsprechend ermittelte man die Mitteltemperaturen der Luft, damit die Wassertemperatur auf die gegebenen vierundzwanzig Stunden zu berechnen.

Solche Berechnung gibt nur annähernde Werte der Wassertemperatur. Ausserdem ist diese Berechnung nur für den Herbst brauchbar. (In unseren Breiten für : September, October und November). In die Sommerperiode ist die Wassertemperatur der nicht tiefen Stauseen als Regel höher als die Mitteltemperatur der Luft.

Um den Berechnungsfehler der Wassertemperatur für die Periode vor dem Eisstand zu verringern, haben wir die Berechnungen ein Monat lang vor dem erwarteten Eisstandtermin durchgeführt. In diesem Falle nähert die Berechnung die Wassertemperaturgrösse zu ihrer wahrer Grösse. (wenn die Berechnungsformel, ebenso wie die Berechnung des Wärmeumsatzes, den Abkühlungsgang naturgemäss wiederspiegeln.)

Der Wärmeumsatz einer Wassermasse wurde auf Grund meteorologischer Standartbeobachtungen für jeden Termin besonder berechnet. Durch ihre Summierung und Multiplikation auf 360 (bei 4 Terminbeobachtungen) bekömmen wir den Wärmeumsatz während vierundzwanzig Stunden. Aus einer grossen Anzahl der Faktoren des Wärmeumsatzes sind vier Faktoren am wichtigsten : die Wärmezufuhr der summaren Sonnenstrahlung ($\sigma_{\bigcirc}$), die Wärmeabgabe durch effektive Ausstrahlung (σ_i), der Wärmeumsatz als Folge der Verdunstung und Wasserdampfkondensation (σ_u) und der konvektive Wärmeumsatz des Wassers mit der Luft (σ_k). Die Zufuhr der globalen Sonnenstrahlung ($\sigma_{\bigcirc}$) wurde mittels der Tabelle von Ukraintzew [2] für den Herbst und für den Frühling bestimmt. Diese Tabelle ergibt einen Mittelwert der Radiation während vierundzwanzig Stunden bei dem wolkenlosen Himmel. Der Wolkeneinfluss wurde durch die Korrektureneinleitung berücksichtigt. Die Korrekturgrössen wurden mittels der aktinographischer Beobachtungsangaben des Kuibyschew Observatoriums and des Michelson observatoriums in Moscau vom Autor festgestellt. (Zeichn, 4).



Zeichn. 4 — Die Beziehung zwischen Werte der globalen Sonnenstrahlung bei trüben und bei wolkenlosen Himmel.
a) 1 — Oktober, 2 — November, 3 — Dezember,
b) 1 — März, 2 — April, 3 — Mai.

Die Zeichn. 4 ist nach den Angaben der Mittelwerte der globalen Sonnenstrehlung und der Bewölkung für eine Dekade zusammengestellt. Dabei wird der kleinere Einfluss von oberen und mittleren Bewölkung auf die Radiation mit Hilfe der Formel (5) berücksichtigt :

$$\text{Fad. } N = N_H + \frac{N_0 - N_H}{2}; \qquad (5)$$

wo N_0 die gesamte Bewölkung in Gr.
N_H die niedere Bewölkung in Grad.

Die Zeichnung 4 zeigt, dass die Genauigkeit der Dekadensummen von der Sonnenstrahlung nicht gross ist.

Für einzelne 24 Stunden sind die Relativfehler der Berechnung noch bedeutender. Da die Berechnung für den Herbst und Frühlung für die Periode mehr als ein Monat lang erfüllt wurde, verringern sich die Relativfehler bemerkbar.

Albedo der Wasserfläche ist nach der Tabelle von Siwkow ([3]) bestimmt. Die effektive Wasserstrahlung (σ_i) wurde anfangs nach der Formel von Brunt ([1]) mit den Koeffizienten von Kusmin ([5]) bestimmt :

$$\sigma_i = \sigma T^4_{\text{wassers}} - A\sigma T^4_{\text{luft}} (1 + CN); \qquad (6)$$

wo T_{Wasser} und T_{Luft} -absolute Wasser und Lufttemperaturen sind, (die letztere auf der Höhe von 2 m.)

σ ist konstante Grösse der Strahlung 8.26×10^{-11} Kal./cm² min. $A = 0{,}65 + 0{,}051 \sqrt{}$ fa;

fa ist absolute Feuchtigkeit in mm;

N ist die Bewölkung in Grad; (besonder für niedere und zusammen für obere und mittlere Bewölkung).

C ist der Koeffizient, der den Einfluss verschiedener Bewölkung auf die Gegenstrahlung bestimmt. Er wurde von Kusmin als 0,12 für die obere und mittlere Bewölkung und als 0,24 für die niedere Bewölkung angenommen.

Nach dem Erscheinen der Monographien von Brasslawski und Wikulina ([6]) wurde die Strahlungsberechnung nach der Formel der letzteren erfüllt.

$$\sigma_i = 0{,}95 \left\{ \sigma T^4_{\text{wassers}} - \sigma T^4_{\text{luft}} \left[\frac{C_H N_H}{10} + \frac{C_{b+c}(Nb + Nc)}{10} + A\left(1 - \frac{N_0}{10}\right)\right]\right\} ; \quad (7)$$

wo $C_H\ C_{b+c}$ sind Die Koeffizienten für niedere, obere und mitllere Bewölkung.
$N_H\ N_B\ N_C$ sind niedere, obere und mittlere Bewölkung in Grad;
N_0 ist Gesammtbedeckung des Himmels;
Im übrigen ist alles ebenso wie bei Brunt.

Die Koeffizientengrössen unterscheiden sich in dieser Formel von den oben erwähnten. Für den Herbst sind die Verschiedenheiten bei der Strahlungsberechnungen nach diesen Formeln gering.

Für den Frühling gibt die Formel von Brunt mit den Korrekturen von Kusmin viel wenigere Strahlungsgrößen, besonders in die Periode hoher Lufttemperaturen.

Der Wärmeumsatz durch die Verdunstung (σ_u) wurde nach der Verdunstungsformel von Saikow ([7]) berechnet und von uns in $\frac{\text{Kal.}}{(\text{cm}^2)\ \text{min.}}$ umgerechnet.

$$\sigma_u = (0{,}0062 + 0{,}0048\ W_{200})\ (fb - fa)\ \frac{\text{Kal}}{\text{cm}^2\ \text{min}} ; \quad (8)$$

wo W_{200} ist die Windgeschwindigkeit in m/sec auf der Höhe von 200 cm. über dem Wasser;

f_b und f_a sind entsprechend sättigende Feuchtigkeit bei der beobachteten Wassertemperatur und absolute Luftfeuchtigkeit über dem Wasserbehälter auf der Höhe von 200 cm. (über dem Wasser);

Die Feuchtigkeit ist nach den Verdunstungsangaben der Wasserbehälter-Verdunstungsmesser nach den Beobachtungen über die Luftfeuchtigkeit und Windgeswindigkeit unmittelbar in der Nähe der Wasserbehälter ermittelt.

Wir haben nur den Koeffizient der Windgeschwindigkeit von 0,72 bis 0,78 vergrösst. Wir sind der Meinung, dass auch in dieser Art die Formel den Wärmeumsatz durch Wasserverdunstung bei starkem Wind verkleinert.

Der konvektive Wärmeumsatz (σ_x) wurde nach der analogischen Formel berechnet. Die Koeffizienten zu dieser Formel wurden von Saikov([7]) auf Grund der bekannten Stellung von Bowen ([8]) ermittelt.

$$\sigma_x = (0{,}0041 + 0{,}003\ W_{200})\ (T - t)\ \frac{\text{Kal}}{\text{cm}^2\ \text{min}} ; \quad (9)$$

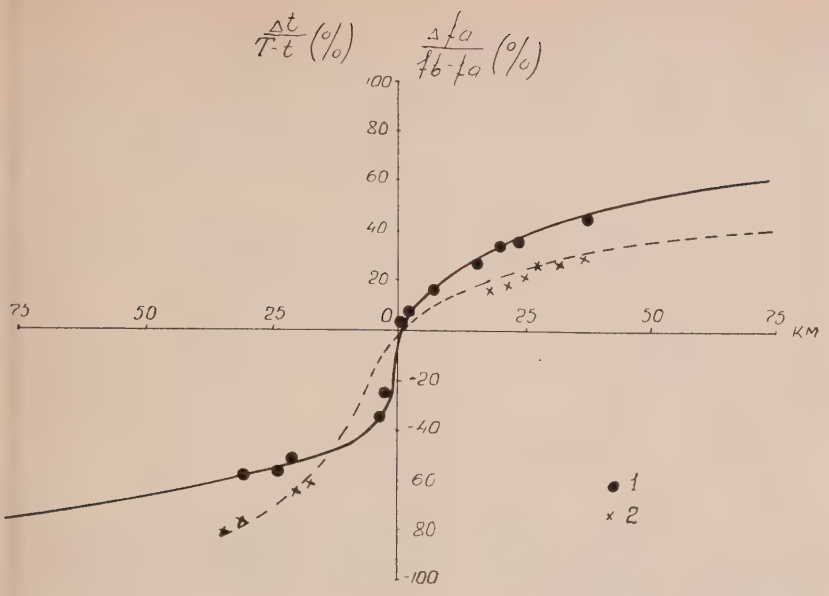
wo T ist die Wasseroberflächentemperatur;
t ist die Lufttemperatur in der Höhe von 2 m über der Wasseroberfläche;
w_{200} ist die Windgeschwindigkeit in m/sec in der Höhe von 200 cm.

Wie es schon bekannt ist, übt der Wasserbehälter auf die Feuchtigkeit und Temperatur der Luft einen Einfluss aus.

Darum ist die Anwendung der oben erwähnten Formeln des Wärmeumsatzes durch Verdunstung und Konvektion zu den Materialen der Standartbeobachtungen von zu Lande liegenden meteorologischen Stationen ohne *Korerktureinleitung* nicht möglich. Diese *Korrekturen*berücksichtigen den Einfluss eines Wasserbehälters auf die Feuchtigkeit und Temperatur der über ihm liegenden Luft.

Die Grösse dieser *Korrekturen* wurde nach den Forschungsmaterialen des Zentralinstituts für Prognosen von Konowodow ([9]) bestimmt.

Auf der Zeichnung 5 ist der Einfluss der Wasserbehältersbreite auf die Feuchtigkeit und Temperatur der über dem Wasserbehälter liegenden Luft gezeigt. Es stellte sich heraus, dass der Einfluss eines Wasserbehälters, also die *Korrekturen*,



Zeichn. 5 — Die Veränderung der Temperatur und Feuchtigkeit der Luft über dem Wasserbehälter im Zusammenhang mit seiner Breite.
1 — Lufttemperatur,
2 — Luftfeuchtigkeit.

erschieden sein können, wenn die Lufttemperatur höher oder tiefer als die Wasser-emperatur ist, und die absolute Feuchtigkeit höher oder niedriger als die die Luft bei der beobachteten Wassertemperatur sättigende Feuchtigkeitgröße ist. Je grösser die Breite eines Wasserbehälters ist, desto grösser die *Korrektur* wird, anfangs — schneller, dann — langsamer.

In die Formeln von Saikow (7) für die Wärmeumsatzberechnung durch Verdunstung und Konvektion ist die Windgeschwindigkeit auf der Höhe von 2 m über dem Wasser eingeschlossen. Anstatt der Windgeschwindigkeit in dieser Formel wurde die Windgeschwindigkeit in der Höhe eines Flügers der meteorologischen Station (in der Höhe von 9-11 m über dem Boden) gestellt. Solches Ersetzen ist darum möglich weil die Windgeschwindigkeit über dem Wasserbehälter auf 1,2 Mal grösser als über der dem Windeinfluss untergeworfenen zu Lande liegenden meteorologischen Station ist, und zwischen der Windgeschwindigkeit in der Höhe eines Flügers und der Geschwindigkeit in Höhe von 2 m über dem Boden fast gleiche Geschwindigkeitsverhältnisse stattfinden.

Die Berechnungsgenauigkeit der mittleren Wassertemperatur nach dem oben beschriebenen System wurde durch die Gegenüberstellung der ermittelten Temperatur der beobachteten mittleren Wassertemperatur auf der Pirogowski Flußstrecke des Klasminskoje Stausees geprüft. (H = 9,2 — 9,3 m).

Mit absicht der Nachprüfung wurden einmal in der Woche im Herbst vom Jahre 1952 bis 1955 Herumfahren der Flußstrecke und oftmalige Temperaturmessungen des Oberflächenwassers und der Flußstreckentiefe durchgeführt. Diese Messungen haben die Möglichkeit mittlere Wassertemperatur einer Flußstrecke zu bestimmen. Die meteorologischen Angaben für die Berechnung des Wärmeumsatzes wurden im

Aerologischen Zentralobservatorium, die sich in iener Entfernung von 12 km zum Westen von der Pirogowski Flußstrecke befindet, genommen.

Im Durchschnitt fallen die berechneten und die beobachteten Temperaturänderungen des Wassers zusammen. Aber für einzelne Periode waren die Unterschiede bedeutend : von + 0°,40 bis- 0°,57. Das wird von uns als Folge einer schlechten Repräsentativität der Flügeranzeigen des Observatoriums für die Bedingungen eines Wasserbehälters erklärt.

Wir interessieren uns für die Zufrierensdata eines Wasserbehälters. Zum Moment des Eisstandes ist die Mitteltemperatur des Wassers gewöhnlich höher als 0°. Die Temperaturgrösse hängt von den Wetterverhältnissen, von der Tiefe und von der Schützung vom Windeinfluss eines Wasserbehälters ab.

Z.B. Pirogowski Flußstrecke, die die mittlere Tiefe 9,2 — 9,3 m und die Breite 1 km hat, bedeckte sich mit Eis in den Jahren 1953-1956 bei der Mitteltemperatur des Wassers 0,3° + 0,5°.

Mehr seichte und breite Iwanjkowskoje Stausee (H = 7 — 9 m, B = 5 — 6 Km) bedeckte sich mit Eis bei folgenden Mitteltemperaturen des Wassers : im Jahre 1938 -bei 0°,15; 1939 -bei 0°,27 und 1940 -bei 0°,12.

Bei einem stillen frostigen Wetter erfolgt das Eisstand bei relativ hohen Mitteltemperaturen des Wassers; bei einem windigen Wetter mit einem schwachen Frost aber ist die Überkühlung des Wassers auf die ganze Tiefe und die Eisbildung innerhalb des Wassers möglich, wie es schon einige Male in Iwankowskoje Stausee vor sich ging.

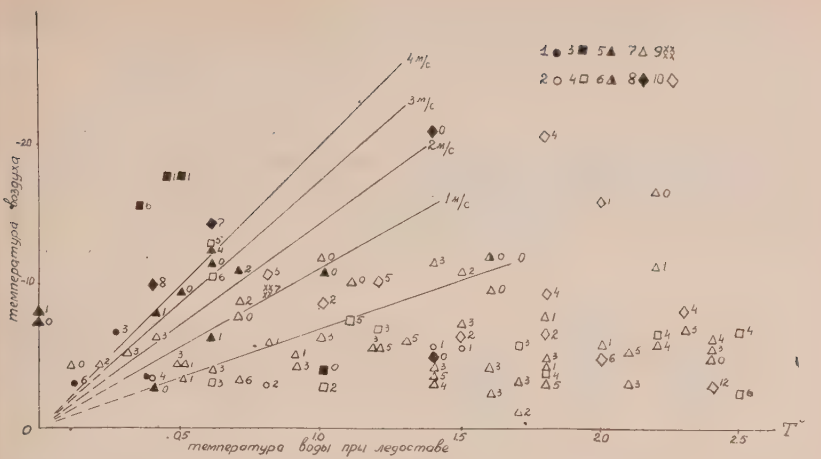
Es wurde die Gegenüberstellung der Eisstandtermine einer Reihe von Wasserbehälter mit den für diese Wasserbehälter ermittelten Data der Wasserabkühlung bis 0° erfüllt. Diese Gegenüberstellung hat gezeigt, dass mehr tiefe (H = 7,5 — 9,2 m) und schmaler (B∼2 km) Pereborski Bucht des Rybinskoje Stausee sich mit Eis auf 1-2 Tage früher bedeckt, als man nach der Berechnung den Termin der Wasserabkühlung bis 0° bekommt. Zu gleicher Zeit erfrieren die breite, aber nicht tiefe Räume der Rybinskoje (bei dem Insel Mologa) und Tsimljanskoje Stauseen (höher Krassnaojrskoje) und die nicht tiefe Pskowskoje See (bei den Insel Salita und Lissje) besonders oft nach dem Temperaturübergang über 0°; in einigen Fällen aber viele Tage später.

Für solche Wasserbehälter fällt der Übergang der Wassertemperatur über 0°, als Regel, mit dem Anfang des Eisganges oder mit der Grundeis zusammen.

Das ist offensichtlich, dass für das Eisstandeintreten bei diesen Verhältnissen eine Eissammlung und sein Zusammenfrieren mit den Ufern notwendig sind, worauf Zeit und eine fortwährende Wärmeabgabe nötig sind. Wenn aber nach dem Temperaturübergang über 0° in kurzer Zeit eine Temperatursteigerung auftritt, dann beginnt das Eisstandeintreten bei dem nächsten Temperaturrückgang, der manchmal viele Tage später auftritt. Gerade in diesen Jahren entsteht eine grosse Differenz in den Übergangsterminen der Wassertemperatur über 0° und in den Daten des Eisstandes.

Man muss also die Eisstandtermine auf den Stauseen mit der Berechnung der Wassertemperatur, Wetterverhältnisse, der Grösse und der Tiefe der Wasserbehälters bestimmen. Als Beispiel einer solchen Abhängigkeit für die Bestimmung der Eisstandtermine wird die Zeichnung 6 angeführt. Diese Zeichnung ermöglicht die Eisstand termine auf den Stauseen mit einer Mitteltiefe 7-10 m zu bestimmen.

Bei der Zusammenstellung der Zeichnung 6 wurde eine grosse Anzahl der berechneten und relativ nicht grosse Anzahl der beobachteten Mitteltemperaturen des Wassers von Iwanjkowskoje, Rybinskoje, Klasminskoje und Tsimljanskoje Stauseen ausgenutzt. Ausserdem wurden alle auf den Meßstellen beobachteten Eisstandtermine



Zeichn. 6 — Wassertemperatur, Lufttemperatur, Windgeschwindigkeit und Eiserscheinungen auf den Stauseen.
Iwanjkowskoje Stausee, die Mitteltiefe 7,6—9,5 m.
1 — Eisstand, 2 — rein oder Salo (Die an der Oberfläche fließende Zusammen gefrorene Eiskristalle) Pirogowski Flussstrecke des Stausees Kljasminskoje, die Mitteltiefe 7,3—8,2 m.
3 — Eisstand, 4 — rein oder Salo.
Die Bucht Pereborski des Stausees Rybinskoje, die Mitteltiefe 7,3—8,6 m.
5 — durchgehender Eisstand, 6 — Eisstand stellenweise, 7 — rein oder Salo.
Der Stausee Tsimljanskoje, die Mitteltiefe 7,7—9,5 m.
8 — Eisstand, 9 — Eisgang, 10 — rein oder Salo.
Neben den Punkten — die Windgeschwindigkeit auf der meteorologischen Station.

und die Angaben der Lufttemperatur und Windgeschwindigkeit der nächstliegenden meteorologischen Stationen benutzt.

Man muss anmerken, dass die Angaben gewöhnlicher Temperaturbeobachtungen bei den Ufern der Stauseen und an den Meßstellen für die *Schätzung* des thermischen Zustandes solcher Wasserbehälter wegen ihrer *Unreprasentativität* nicht brauchbar sind. Die Mitteltemperatur des Wassers für wenig durchfliessende Wasserbehälter kann man mittels vielmaliger und ausführlicher Beobachtungen in verschiedenen Stellen und Tiefen bekommen.

Auf Grund des beschriebenen Systems wurden die Eisstandtermine für eine Reihe der neugeschaffenen Stauseen bestimmt.

In der Tabelle 1 sind die berechnete jährliche Eisstandtermine für ein tiefes Nebendammgebiet und für einen oberen nicht tiefen Teil des Kuibyschewskoje Stausees während einer Jahresreihe im Vergleich zu den in jene Zeit auf dem Fluss beobachteten Eiserscheinungen gegeben.

Man kan sehen, dass der Eisstand auf einem Stausee stark ausgedehnt in den Zeit wird. Zuerst beginnen obere nicht tiefe Gebiete (H = 4 m) zu erfrieren, und zwar in die dem Eisgang auf den Flussen Wolga und Kama in natürlichen Zustand nahe Termine und in bedeutend frühere Termine im Vergleich zu den Eisstandterminen auf einem Fluss.

Das tiefe Nebendammgebiet (H = 17 m) würde sich in den betrachteten Jahren viel später als die oberen Gebieten und in gewöhnlich spätere Eisstandtermine, als

die an der Wolga in diesem *Gebiet*, mit Eis bedecken. Man muss unterstreichen, dass man dieselben Ergebnisse bei der Bestimmung der Mitteldata ermittelt.

TABELLE 1

Die berechneten Eisstandtermine für den Stausee Kuibyschewskoje.

Nebendammgebiet H = 17 m.

	Berechnungsanfang	Anfangstemperatur des Wassers	Datum der Wasserabkühlung bis 0°	Eisstanddatum des Wasserbehälters	Anfangsdatum des Eisganges an der Wolga bei Morkwaschi	Anfangsdatum des Eistandes an der Wolga bei Morkwaschi
1945	12/X	11,2	29/XI	25/XI	9/XI	21/XI
46	17/X	10,0	10/XII	5/XII	29/X	26/XI
47	24/X	9,2	14/XII	11/XII	23/XI	15/XII
48	24/X	8,7	14/XII	8/XII	16/XI	8/XII
49	29/X	7,1	11/XII	12/XII	15/XI	4/XII
50	20/X	11,3	15/XII	12/XII	22/XI	12/XII

Ausläufer Kamsky. H = 4 m.

	Berechnungsanfang	Anfangstemperatur des Wassers	Datum der Wasserabkühlung bis 0°	Eisstanddatum des Wasserbehälters	Anfangsdatum des Eisganges an der Kama bei Tschistopol.	Anfangsdatum des Eisstandes an der Kama bei Tschistopol.
1945	11/IX	14,4	25/X-6/XI[1]	9/XI	24/X	17/XI
46	17/IX	14,3	24/X	26/XI	25/X	24/XI
47	26/IX	10,6	5/XI-22/XI[1]	25/XI	8/XI-24/XI	8/XII
48	1/X	10,2	14/XI	15/XI	14/XI	7/XII
49	26/IX	10,8	12/XI	12/XI	15/XI	27/XI
50	26/IX	11,8	3/XI	5/XI	3/XI	24/XI

(1) Die Temperatursteigerung der Luft am Tage der Wasserabkühlung bis 0°; infolgedessen begann die Temperatursteigerung des Wassers bis 0° und der endgültige Temperaturübergang über 0° erfolgte im Datum, das im Nenner vorgeschrieben ist.

Die Bestimmung der Jährlichen Reinigungstermine.

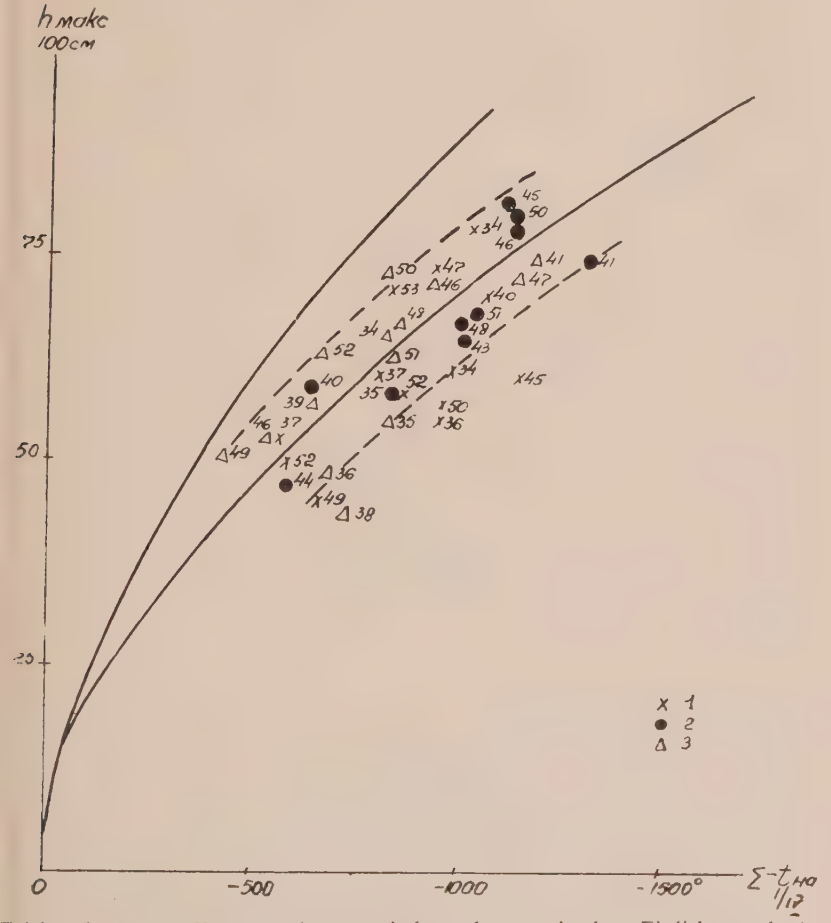
Die jährlichen Reinigungstermine der neugeschaffenen Stauseen bei den meteorologischen Verhältnissen vorigen Jahre wurden nach der Wärmebilanzmethode festgestellt.

Die Reinigungstermine wurden durch die Gegenüberstellung der für das Tauen einer Schnee und Eisdecke eines Wasserbehälters notwendigen Wärme der Wärmebilanz dieses Wasserbehälters mit dem umliegenden Raum ermittelt.

$$80\,(\rho_\Lambda\, h_\Lambda + \rho_{c\Lambda}\, h_{c\Lambda} + \rho_c\, h_c) = \sigma\tau; \qquad (10)$$

wo h_Λ $h_{c\Lambda}$ h_c $b_{zw.}$ die Dicken eines gewöhnlichen Eises, eines *Schneeeises** und eines auf dem Eis liegenden Schnees sind und

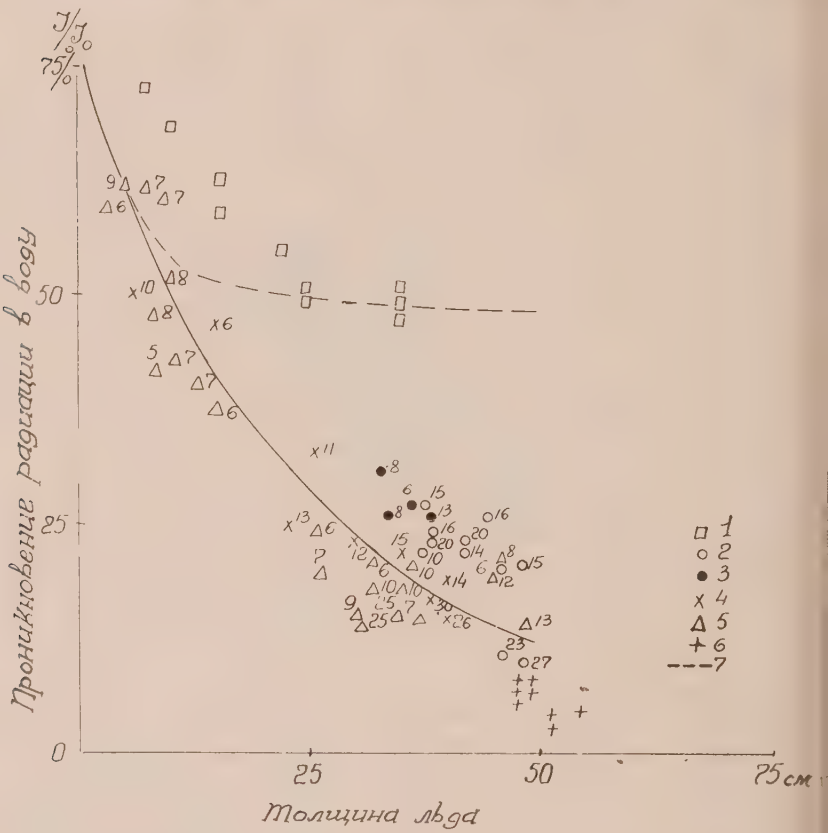
ρ_Λ; $\rho_{c\Lambda}$; ρ_c $b_{zw.}$ ihre Dicke.



Zeichn. 7 — Der Zusammenhang zwischen der maximalen Eisdicke und der Tagessumme negativer Luftmitteltemperaturen für die Eisstandperiode.
1 — auf dem Fluss Wolga bei Dubowka.
2 — auf dem Stausee Rybinskoje.
3 — auf dem See Ilmen.

* Das Eis welcher sich bei dem Gefrieren der mit dem Wasser ersättigten Schnees bildet.

Solche Detaillierung bei der Bestimmung des Eismaterialbestandes ist nur bei den Beobachtungen auf den existierenden Wasserbehälteren möglich. Für die zukünftigen Stauseen ist nur eine annähernde Berechnung dieser Grössen möglich. Für diese Zwecke wurde die Abhängigkeit (7) zwischen der maximalen Eisdicke vor dem Tauenanfang und der Summe negativer Mitteltemperaturen der Luft in vierundzwanzig Stunden für die Eisstandperiode festgestellt.Diese Abhängigkeit wurde nach den Angaben über die mittlere Maximaldicke von Eis auf den Messstellen von Rybinskoje Stausee, Ilmen See und anderen aufgestellt. Diese Abhängigkeit ermöglicht die Maximaldicke einer Eisdecke mit Genauigkeit $\pm$ 15 % der zu bestimmenden Grösse zu ermitteln. Heutzutage werden die Forschungsarbeiten mit absicht der Präzisierung



Zeichn. 8 — Durchdringen der Sonnenstrahlung durch Eisdecke ins Wasser im Zusammenhang mit der Eisdicke.
1 — die Angaben von Kalitin für die gerade Radiation bei Schnee- und Eisschneefehlen.
2 — die Angaben von Arssenjewa ([10]) für den zweite See in Oserki.
3 — die Angaben für den Stausee Kljasminskoje für das Jahr 1954.
4 — die Angaben für den Stausee Kljasminskoje für das Jahr 1955.
5 — die Angaben für den Stausee Kljasminskoje für das Jahr 1956.
6 — die Angaben von Kalitin und Arssenjewa ([10]) bei dem Vorhandensein des Schneeeises.
7 — die Angaben von O. Devik (1) für den Fall des idealenreinen Eises. Bei den Punkten ist die Albedogrösse einer Eisdecke eingeschrieben.

dieser Abhängigkeit durchgeführt. Das wurde hauptsächlich mit Hilfe der Statistik des Schneeeises, dieser gewöhnlichen Erscheinung auf den Wasserbehältern des Nordens und des Zentrums Europäischen Teils der UdSSR ermöglicht. Der Eisdeckenwärmeumsatz in die Frühlingsperiode wurde mit der Berechnung von vier Hauptbestandteile — der globale Sonnenstrahlung, der effektiven Ausstrahlung, Verdunstungo der Kondensation und des konvektiven Wärmeumsatzes ermittelt — ebenso wie bei den Berechnungen der Abkühlung eines Wasserbehälters, aber mit der Berechnung einiger Besonderheiten des Tauenprozesses einer Eisdecke.

So ist für den Frühling das Durchdringen der Sonnenstrahlung unter das Eis (ins Wasser) sehr wichtig, da ein Strahlungsanteil auf die Wasserwärmung verbraucht wird. Im Frühling schwankt Albedo der Eisdecke stark. Eine grosse Bedeutung hat die Wärmezufuhr zur Eisdecke von unten (vom Wasser), welche von der Eistauen von unten hervorgerufen ist.

Für die Klärung dieser Momente wurden in den Jahren 1954—1956 Speziallbeobachtungen auf der Pirogowski Flusstrecke des Klasminskoje Stausees durchgeführt. Auf der Zeichnung 8 ist das Durchdringen der Sonnenstrahlung durch Eisdecke verschiedener Dicke im Frühling beim Schneefehlen auf der Eisoberfläche gezeigt. Beim Vorhandensein von Schnee auf dem Eis dringt die Radiation ins Wasser fast gar nicht durch und spiegelt sich hauptsächlich wieder. Eine grosse Bedeutung für das Radiationsdurchdringen ins Wasser hat das *Schneeeis*. da das letzte die Radiation stark aufsaugt. In Wirklichkeit dringt eine bemerkbare Menge der Sonnenstrahlung ins Wasser durch und erwärmt es nur nach dem Schnee-und *Schneeeistauen*. Aber die Menge der ins Wasser durchdringenden Radiation für die Tauenperiode eines Wasserbehälters ist höchstens auffällig. Das wird weiter gezeigt.

Die Eisdeckenalbedo ist im Laufe des Frühlings sehr verschieden. Sie hängt vom Vorhandensein von Schnee, *Schneeeis*, Wasser auf dem Eis und vom Zerstörungsprozes einer Eisdecke ab. Bei einem neugefallenenen trockenen Schnee ist die Albedo gross (70—80 %). Nach der Schneefirnisation und nach dem Erscheinen ihm Schnee einer nicht grossen Menge von Wasser senkt Albedo bis 60 %. (aber nicht weniger als 50 %). Im Schneetauenprozess sättigt sich der Schnee mit Wasser, und das letzte erscheint auf der Eisdeckenoberfläche. Die Albedo des Wassers und des mit Wasser gesättigten Schnees, unter welchem noch *Schneeeis* liegt, ist nicht gross (bis 20 %).

Nach dem Wasserabfluss unter das Eis nimmt die Albedo wieder zu und erreicht die Grösse von 30 %. Das ist dadurch zu erklären, dass auf der Oberfläche *Schneeeis* und Schneereste liegen. Ferner hängt die Albedo eines schmelzenden *Schneeeises* von seinem Sättigungsgrad mit Wasser ab: beim Vorhandensein einer grossen Wassermenge nimmt Albedo bis 20 % ab, und bei dem Wasserabfluss unter das Eis nimmt Albedo bis 30 % zu.

Sehr stark verringert sich Albedo (bis 10—15 %) und vergrösst sich das Radiationdurchdringen ins Wasser bei dem Hinausgehen Kristalleises auf die Oberfläche. In diese Periode bemerkt man, dass das Eis dunkel wird. Später nimmt Albedo langsam ab und erreicht die für die Wasseroberfläche charakteristischen Grössen 7—9 %; und das Radiationdurchdringen ins Wasser nimmt zu.

Bei der Bildung einer kompakten Schneedecke auf dem Eis im Frühling kehren Albedo und Radiationdurchdringen mit einem Sprung zu den für diese Verhältnisse gewöhnlichen Grössen zurück: Albedo beträgt 60—70 % und das Radiationdurchdringen wird fast nicht bemerkbar.

In der Tabelle 2 sind die Grössen von Albedo einer Eisdecke und das Radiationdurchdringen ins Wasser für die ganze Tauenperiode auf dem Klasminskoje Stausee (nach den Angaben des Autors) und auf dem See Oserki bei Leningrad im Jahre 1941. (Nach den Angaben von Arssenjewa ([10])) gegeben.

TABELLE 2

Das Jahr	Die Albedo %	Das Durchdringen der Radiation ins Wasser %	Die Albedo + Durchdringen der Radiation %
1941	28	16	44
1954	22	23	45
1955	26	18	44
1956	20	22	42

Für langwierige Frühlinge mit oftigen Schneefällen (1941) und für die Eisdecke mit einer dicken *Schneeeisschicht* (1955) war Albedo grösser als in anderen Jahren. *Das Radiationdurchdringen ins Wasser In solchen Jahren* war verkleinert. In der Summe betrugen Albedo und Radiationdurchdringen etwa 44 %, und diese Summe war von Jahr zu Jahr stabil.

Bei den Berechnungen des Eisdeckenabtauens haben wir angenommen, dass die Eisdecke 55 % auf den Eis fallender globalen Sonnenstrahlung einsaugt. Vielleicht ist diese Grösse verkleinert, weil ein Teil der mit oberen Wasserschichten eingesaugten Sonnenstrahlung auf das Eistauen von unten verbraucht wird.

Der Wärmestrom vom Wasser zur Niedergrenze einer Eisdecke wurde von uns auf den speziellen Anlagen studiert. Dafür wurde die Eisdecke mit einer dicken Schneeschicht von oben belastet, oder wurde die obere Eisschicht beseitigt; infolgedessen trat das Wasser aufs Eis und überschwemmte es von oben. Auf solche weise wurde das Eis mittels eine Wasserschicht vom äusseren Kälteeinfluss isoliert. Die Eistemperatur wurde 0°, und der ganze vom Wasser zur Untergrenze der überschwemmten Eisdecke gerichtete Wärmestrom rief das Tauen hervor. Das Tauen wurde mit Hilfe der genauen Latte von einem ständigen Pegel der in den Eis eingefrorenen Pfalen bestimmt.

Auf der Pirogowski Flussstrecke wurden die Grössen einer Wärmezufuhr zur unteren Eisgrenze für die ganze Periode des Eistauens bestimmt.

Im Jahre 1955 betrug diese Grösse 770 Kal/cm²; im Jahre 1956—940 Kal/cm².

Bei den Berechnungen der effektiven Ausstrahlung, der Verdunstung und des konvektiven Wärmeumsatzes ist es notwendig die Temperatur der Eisdeckenoberfläche zu haben. Für solche Temperatur wurde 0° bei einer positiver Lufttemperatur angenommen, d.h. dass man vermutete, dass bei diesen Verhältnissen die Eisdecke von der Oberfläche taut. Bei einer negativen Lufttemperatur vermutet man eine gleiche Temperatur für die Eisoberfläche und für die Luft. Diese Annäherung in den Temperaturen ist sichtbar, da die Temperatur einer Eisoberfläche von den Gesammtbedingungen des Wärmeumsatzes abhängig ist, wie das O. Devik (1) und spezielle Beobachtungen des Autors gezeigt haben. Aber die infolgedessen entstehenden Fehler sind nicht gross, darum kann man sie vernachlässigen.

Die Brauchbarkeit des oben beschriebenen Systems wurde durch Erfahrungsberechnungen der Reinigungstermine von Rybinskoje, Tsimljanskoje und Klasminskoje Stauseen und Tschudskoje, Pskowskoje und Ilmen Seen für eine Jahres reihe geprüft. Die meteorologischen Angaben für diese Berechnungen wurden von den nächsten zu Lande liegenden meteorologischen Stationen genommen. Der Wärmeumsatz wurde vom Tage des Frühlingsüberganges der Lufttemperatur über 0° nach den viermaligen meteorologischen Terminbeobachtungen berechnet. Die vierundzwanizg Stunden, *von denen* an die nach der Wärmebilanz berechnete summare Wärmezufuhr

zur Eisdecke eine für Eistauen notwendige Wärme zu überschreiten begann, wurden für das Reinigungsdatum gehalten. Die wirklichen Reinigungstermine wurden nach den Beobachtungsangaben der Messstellen festgestellt.

Die Kontrollberechnungen haben gezeigt, dass die berechneten Reinigungstermine für breite Teile der Stauseen und der Seen auf 1—6 Tage früher der beobachteten Data einer vollen Reinigung waren. Die Berechnung gibt dem Schiffahrtanfang nähere Data, weil die vom Wind zum Ufer gedrückten Eisreste gewöhnlich hier bis Ende tauen. Zu dieser Zeit sind schiffbare, vom Ufer entfernte Trassen eisfrei, und auf den letzten schon die Schiffahrt efolgt. Die berechneten Reinigungstermine für solche schmale Wasserbehälter wie Pereborski Bucht (B$\sim$2,0 km) und Pirogowski Flussstrecke des Klasminskoje Stausees (B$\sim$1 km) gingen auseinander, und die Differenz war $\pm$ 3 Tage.

Nach der oben erwähnten Methodik wurden die Reinigungstermine für einige sich zu spät reinigende breite Teile einer Reihe neugeschaffener Stauseen bestimmt In der Tabelle 3 sind diese Termine in den Jahren 1945—1950 für den Kuibyschewskoje Stausee im Bezirk seiner Erweiterung bei Uljanowsk im Vergleich mit den Anfangsterminen des Eisganges und der Reinigung der Wolga, gegeben.

TABELLE 3

Die berechneten Reinigungstermine des Stausees Kuibyschewskoje bei Uljanowsk im Vergleich mit den Daten des Eisganganfanges und der Reinigung vom Eis.

Jahr	Summe Lufttemperaturen für die Eisstandperiode	Für Eistauen notwendige Wärme cal/cm²	Berechnete Termin der Reinigung	Termine für Wolga bei Uljanowsk		Differenz in den Terminen für Wolga und für Stausee
				Eisgang	Reinigung	
1945	—1726	7600	11/V	23/IV	1/V	10
1946	—1403	6800	2/V	24/IV	1/V	1
1947	—1492	7200	27/IV	5/IV	17/IV	10
1948	—1136	6000	2/V	18/IV	29/IV	3
1949	—1032	5760	1/V	16/IV	29/IV	2
1950	—1500	7200	28/IV	16/IV	24/IV	4

Die Reinigung des Kuibyschewskoje Stausees wird später erfolgen, als die an der Wolga. Im Falle eines schnellen Frühlings nähren sich die Termine, und im Falle eines Temperaturrückganges nach der Flussreinigung erfolgt die letzte mit einer Verspätung bis 10 Tage. (In den Jahren 1945—1947).

Diese Berechnungen gaben die Möglichkeit das Regime der Eisstand und Reinigungstermine für einige neugeschaffenen Stauseen auf den Flüssen der USSR zu charakterisieren und eine Reihe der für die Ausarbeitung der Methodik für die langfristige Vorhersage dieser Erscheinungen notwendiger Termine zu ermitteln. Die vorgeschlagene Berechnungsmethodik kann auch für die Bestimmung der Eisstand- und Reinigungstermine solcher Seen und Stauseen, für welche es keine Eisbeobachtungen gebe, verwendbar sein.

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ÉTUDE SUR LE GEL DU SOL

STÉPHANE HENIN
(*France*)

Résumé

Si l'étude du gel présente un intérêt particulier dans les régions à climat très froid, cette étude présente également une certaine utilité dans les régions à climat tempéré. Non seulement les phénomènes intéressent au plus haut point l'agriculture, mais se présentant avec moins d'intensité, il est possible d'en mieux saisir les détails.

Le sol gèle de deux manières nettement différentes suivant son état. En milieu compact et relativement sec de petits cristaux de glace se forment qui cimentent les éléments du sol. En milieu plus humide, il se forme en outre de véritables lamelles de glace. Les phénomènes de soulèvement sont alors modérés. Dans les terres meubles constituées par des agrégats distincts, il se développe entre les particules de terre des cristaux de glace en aiguilles. En milieu très humide, ces cristaux peuvent se développer d'une manière très considérable entraînant des phénomènes de soulèvement importants.

— Mesure de soulèvement de la terre dans ces différentes conditions.
— Comportement des pierres suivant leur nature.
— Etude de la compression exercée dans le sol.
— Influence sur la structure du sol.

L'étude du gel du sol présente du point de vue pratique un intérêt très général. Il a une action sur les roches qu'il fait éclater, il modifie le sol et détériore les ouvrages d'art, les routes en particulier. Il provoque de graves dégâts sur les cultures qui passent l'hiver en terre. Enfin, il prépare les phénomènes d'érosion tant par le vent que par l'eau. En climat froid ces effets sont considérables.

Nous avons effectué une série de recherches dans la région parisienne et dans le Centre de la France, c'est-à-dire sous des climats où les hivers ne sont généralement pas très rigoureux et où cependant les dégâts dus au gel se manifestent d'une manière notable.

Action du gel sur la distribution de l'eau dans le sol :

Bien que ce ne soit pas toujours là le phénomène le plus visible, l'action du gel provoque toujours une modification dans la distrigution de l'humidité. Celle-ci tend à augmenter et parfois d'une manière considérable dans la zone gelée. Le tableau I fait apparaître ce phénomène. Il s'agit d'un sol de limon contenant 18 % d'argile et 3 % de matière organique d'une part, et d'un sol calciare renfermant 15 % d'argile et 10 % de carbonate de calcium. Ces matériaux ont été placés dans des pots de 40 cm de hauteur, enfouis en terre de manière que leur niveau affleure celui du sol environnant et avant le gel le fond du pot couvert de sable grossier avait été saturé d'eau. On constate une élévation considérable de l'humidité de la couche gelée alors que les couches sous-jacentes présentent une légère diminution d'humidité dans le cas de l'horizon A et une légère augmentation dans le sol calcaire. Toutefois, dans les deux séries d'essais la couche inférieure située un peu au-dessus du plan d'eau est restée à la même humidité. Ceci montre que les variations d'humidité de la couche gelée ont du se faire au moins en partie par remontée de l'eau provenant du plan d'eau de la couche de sable.

TABLEAU 1

Humidité à différentes profondeurs dans deux types de sol

	Horizon A		Sol calcaire	
	non gelé	gelé	non gelé	gelé
Surface	29,9	51,0	24,1	63,3
Courbe sous-jacente	30,6	28,3	24,8	26,6
Mi-profondeur	31,9	30,7	25,6	25,1
Fond	35,3	35,2	27,3	27,3

Le phénomène est toutefois identique en l'absence de plan d'eau et GESLIN (1) avait déjà montré que dans le sol de Versailles en place l'accroissement d'humidité dans la couche gelée s'effectuait aux dépens de l'humidité des couches profondes.

Cette succion qui s'exerce sur l'eau au moment du gel apparaît très nettement quand on mesure le niveau du plan d'eau avant, pendant et après la période froide. MASCHHAUPT (2) avait observé pendant des hivers particulièrement rigoureux que le plan d'eau de ses lysimètres situés à 1 mètre de profondeur subissait des rabattements pouvant atteindre 29 cm pendant les périodes de grand froid. Pendant l'hiver 1956-57, nous avons pu observer à Versailles des abaissements de 30 cm pour des plans d'eau situés à 50 cm et 1 mètre et des températures minima de —7° à —8°. Pour un plan d'eau situé au voisinage de la surface l'abaissement a atteint 60 cm pour les mêmes temperatures.

Cet ensemble de résultats qui confirme ceux obtenus par d'autres auteurs met clairement en évidence un premier aspect du mécanisme du gel qui se traduit par un mouvement de bas en haut de l'eau contenue dans le sol. L'effet est d'autant plus marqué que le plan d'eau est situé près de la surface du sol.

CAUSES DU MOUVEMENT DE L'EAU :

On a cherché la cause de ce phénomène. L'explication parait être la suivante. Le passage de l'eau à l'état de glace équivaut à une dessiccation puisque le liquide n'étant plus susceptible de se mouvoir n'exerce plus sa pression hydrostatique. Il y a donc une tendance au mouvement per ascensum analogue à celui qui se manifeste au cours de l'évaporation. Il faut toutefois remarquer une différence sensible entre l'action du gel et celle de l'évaporation, c'est la vitesse du déplacement de l'eau dans le premier cas. En fait, ce qui parait limiter les mouvements des liquides sous l'influence de la sécheressse c'est la baisse rapide de la conductibilité à mesure que l'humidité décroît. Il semble qu'au contraire lors de l'action du gel, l'humidité ne diminue pas d'une façon sensible et que la conductibilité restant élevée l'eau puisse se déplacer rapidement (KRETSCHMER (3)). Il semble que l'on puisse attribuer l'aspect particulier de l'action du gel au fait que l'eau n'est pas extraite du sol, elle reste sur place sous forme de glace au lieu d'être perdue à l'état de vapeur. Il s'ensuit que les variations d'humidité que l'on observe peuvent rester du même ordre que celles qui se manifestent au début du phénomène d'évaporation, les quantités accumulées sous forme de glace restant du même ordre de grandeur que celles qui sont perdues à l'état de vapeur.

Il faut, enfin, ajouter qu'en présence de plan d'eau à faible profondeur le problème n'a plus de sens car les gros capillaires pouvant se remplir à partir de la nappe d'eau libre, la conductibilité reste élevée.

On a parfois imaginé que le déplacement de l'eau s'effectuait à l'état de vapeur. On peut montrer que c'est par alimentation des cristaux à partir d'eau libre que s'effectue l'accumulation dans certains points du sol (S. Henin, L. Locussol et O. Robichet (4)).

Si l'on dispose une bougie poreuse pleine d'eau en l'enfonçant à la base dans un terrain humide de manière à ce que le liquide reste enfermé à l'intérieur, on constate qu'après une nuit de gel celle-ci est recouverte d'une couche de glace. Si au contraire toute ou partie de la surface poreuse est enduite d'un corps hyrophobe la couche de glace n'apparaît que sur les surfaces qui n'ont pas été enduites. Ainsi, le développement des cristaux de glace parait bien liés à une alimentation directe par la réserve d'eau. D'ailleurs lorsqu'on examine les aiguilles de glace qui naissent à la surface du sol, on constate qu'elles ont englobé de petits fragments de terre ce qui exclue une formation par condensation d'eau.

Effets produits sur le sol :

L'un des effets les plus évidents est le soulèvement du niveau du terrain sous l'influence du gel; l'importance de ce soulèvement est extrêmement variable. Il ne dépend pas nécessairement de l'abaissement de la température, mais plutôt de la façon dont cet abaissement s'établit. Nous reproduisons dans le tableau II les variations de niveau d'une surface de terre meuble et d'une surface de terre tassée. On constate que les effets ne sont pas très différents. Tout au plus les faibles gelées sont-elles un peu plus marquées en sol non tassé. En plus des variations de niveau consécutives aux périodes de gel et de dégel, en observe un affaissement moyen continu de la surface qui est évidemment beaucoup plus prononcé quand le terrain n'était pas tassé au départ.

Il semble que l'on puisse attribuer ces irrégularités à la façon dont s'accumule l'eau gelée.

On peut distinguer deux cas extrêmes (5) : en terre très tassée, il se forme des lames de glace qui s'intercalent par couches à epu près horizontales dans le terrain. Dans les terres meubles, motteuses, il se forme entre les particules de terre des aiguilles de glace qui se développent plus ou moins. Il semble que ce soit dans ces conditions que l'on obtienne les soulèvements les plus importants. En particulier, en sol humide, il est fréquent de constater pour de petits abaissements de température des soulèvements notables de l'ordre de 2 ou 3 cm n'affectant qu'une couche de sol d'une assez faible épaisseur. Ces accidents exercent une influence particulièrement marquée sur les cultures et peuvent provoquer le déchaussement des végétaux, en particulier des céréales d'hiver. On aura une idée de l'effet de soulèvement en examinant l'action d'une série de gels et de dégels sur des piquets de bois (tableau II).

Il semble assez facile d'expliquer cette formation de cristaux en aiguilles. En effet, dans une terre meuble, il y a une certaine discontinuité de la température du sol quand celui-ci est formé d'agrégats bien distincts. Il s'ensuit qu'à un niveau donné les particules de terre sont à une température suffisamment basse pour provoquer la congélation, alors que les particules de terre sous-jacentes sont encore à une température encore un peu supérieure. La glace se forme donc au contact des grains les plus froids et se développe à partir de l'eau extraite des particules les plus chaudes On peut mettre nettement en évidence cet effet de dissymétrie par l'expérience suivante : si l'on enfouit dans une terre humide une surface poreuse, par exemple un fragment de brique de manière qu'un film d'eau vienne affleurer la surface, ce film

TABLEAU II

Influence de facteurs climatiques sur les variations de profondeur de piquets de niveau du sol (moyennes exprimées en cintimètres).

Périodes des relevées	Etat du sol	Profond. de gel (moyen.)	Sol non tassé				Sol tassé			
			Piquets en bois			Niveau du sol	Piquets en bois			Niveau du sol
			15,0	10,0	5,0		15,0	10,0	5,0	
Etat initial	Légèrement humide		0	0	0	0	0	0	0	0
2/11/49-15/12/49	Gelé	1,3	+ 0,25	+ 0,60	+ 0,15	— 0,50	— 0,40	— 0,20	0	+ 0,15
	Dégelé		+ 0,45	+ 0,65	+ 0,20	— 1,15	— 0,25	— 0,10	0	+ 0,20
15/12/49-30/12/49	Gelé	1,0	+ 0,65	+ 1,15	+ 0,05	— 0,95	— 0,60	— 0,15	+ 0,15	+ 0,35
	Dégelé		+ 1,05	+ 1,60	+ 0,25	— 1,45	— 0,15	+ 0,20	+ 0,40	+ 0,25
30/12/49-19/1/50	Gelé	1,0	+ 1,05	+ 1,40	+ 0,15	— 1,15	— 0,30	+ 0,15	+ 0,50	+ 0,25
	Dégelé		+ 1,15	+ 1,55	+ 0,15	— 1,50	— 0,15	+ 0,30	+ 0,55	+ 0,10
28/1/50	Gelé	22,0	+ 1,22	+ 1,74	+ 0,20	+ 0,80	— 0,30	+ 0,30	+ 0,75	+ 2,25
4/2/50-23/2/50	Gelé	0,2	+ 2,20	+ 2,20	+ 0,30	— 1,50	+ 0,95	+ 0,60	+ 0,85	+ 0,05
	Dégelé		+ 2,45	+ 2,25	- 0,35	— 1,85	+ 1,00	— 0,75	- 0,90	— 0,10
23/2/50-3/3/50	Gelé	2,1	+ 3,45	— 2,85	- 0,65	— 0,50	+ 0,95	- 1,25	+ 1,00	- 1,25
	Dégelé		+ 4,10	+ 3,30	+ 0,80	— 1,90	+ 1,85	- 2,25	+ 1,20	— 0,15
3/3/50-15/6/50	Après pluies abondantes		— 5,05	+ 3,40	»	— 2,25	+ 2,50	— 2,70	+ 1,15	— 0,45

NOTE. — Les variations de profondeur des piquets sont mesurées par rapport à la surface du sol aux périodes correspondantes.
Le signe + indique une élévation, le signe — une abaissement.

d'eau se prendra en glace pendant les nuits froides et aura tendance à épaissir, mais formera une couche parallèle à la surface de la brique. Si, au contraire, on dépose quelques petits fragments de terre à la surface du film d'eau, on constate la formation d'un pédoncule de glace de quelques centimètres de hauteur supportant l'agrégat de terre sous l'influence de l'abaissement de température nocturne. Cette expérience nous paraît donc démontrer le rôle de la discontinuité du milieu sur la formation des pédoncules de glace.

On retrouve d'ailleurs cet effet dans le sol lui-même au contact des pierres qui sont enfouies. Il faut toutefois distinguer deux cas : celui qui correspond à notre essai intéresse les matériaux compacts fragment de silex, quartzite etc... on observe qu'il se constitue a leur contact avec le sol des pédoncules de glace comme dans l'expérience que nous avons relatée. Ceci n'est toutefois exact que si la terre est elle-même suffisamment meuble.

Si la pierre ou le gravier sont constitués de matériaux poreux : calcaire, fragment de brique, ils s'entourent de manchons de glace formés par de l'eau extraite de la pierre et du sol, car l'épaisseur de glace est telle qu'il est peu vraisemblable que toute cette eau ait pu être logée dans le fragment de roche consisidéré.

Il est beaucoup plus difficile d'expliquer la formation des lentilles de glace. Certes, dans différents cas, on peut admettre que celles-ci prennent naissance dans une discontinuité du sol, mais dans certains milieux assez compacts on comprend assez difficilement l'existence de ces discontinuités. Il faut, d'ailleurs, souligner que le développement des lentilles de glace n'est pas toujours horizontal et il est fréquent de voir se développer à la surface du sol des séries de fissures rectilignes se raccordant à angle aigü et correspondant à des lamelles développées plus ou moins verticalement

Nous pensons qu'il faut chercher l'explication de ce type de formation dans la dynamique du gel lui-même. Quand le gel se produit sur un milieu homogène le gradient de température étant vertical confère une certaine symétrie au milieu et l'eau tend à venir s'accumuler au niveau de l'isoterme 0^0. Au moment où la glace commence à se former, il y aurait un retard à la pénétration de l'onde de froid favorisant l'accumulation de l'eau, puis après une légère dessiccation des couches inférieures ralentissant la formation de glace l'onde de froid reprendrait sa progression jusqu'à ce que se reconstituent dans le sol les conditions nécessaires à la formation d'une nouvelle couche de glace. Cette explication ne vaut que si le milieu est suffisamment homogène et chaque hétérogénéité pourra se manifester en changeant la forme de la lentille de glace. En ce qui concerne les cristaux verticaux, on pourrait admettre que ceux ci s'amorcent dans les fissures provoquées par le gel des premières couches de terre. Il faut toutefois reconnaître qu'il s'agit là d'hypothèses qui doivent être vérifiées par l'expérience.

Par contre, l'influence de la vitesse de pénétration de l'onde de froidest tout à fait évidente. Si l'on ouvre une fosse dans la terre gelée, après une assez longue période de froid, on constate que dans la couche supérieure du sol il ne s'est développés que de petits cristaux ou de pteites lentilles de glace intimement mélangés à la masse de terre. Au contraire, à la limite de la couche gelée, il tend à se développer soit des lamelles de glace continues, soit des couches de cristaux. L'épaisseur de ceux-ci dépend de la teneur en eau du sol.

On retrouve d'ailleurs expérimentalement le même aspect du phénomène. Si nous reprenons l'expérience de la brique poreuse recouverte d'une lame d'eau et s'il survient une période de gel intense, il ne se forme presque pas de pédoncules de glace parfois même il ne s'en forme pas du tout, l'eau étant prise en masse avant d'avoir eu le temps de se déplacer. Ceci explique également que dans les zones humides il puisse se constituer des cristaux dès glace qui poussent à la surface du sol, alors que le sous-sol n'est pas gelé. Très souvent, dès que l'humidité est un peu moins forte, la couche de cristaux se développe plus à la surface, mais sous une couche de terre de

quelques millimètres. Lorsqu'il existe une mince couche de feuilles reposant sur un sol humide, c'est souvent là que l'on observe les formations de cristaux de glace les plus importantes.

On trouve souvent dans la littérature l'affirmation que le soulèvement du sol est la conséquence de l'accroissement de volume consécutif au passage de l'eau à l'état de glace : c'est certainement inexact. En effet, d'après ce que nous venons d'exposer il est déjà évident que les effets observés sont liés à une nouvelle distribution de l'eau au moment où elle passe à l'état solide. Mais si le soulèvement était seulement du à l'accroissement de volume de l'eau on devrait s'attendre à observer des compressions à l'intérieur du sol. Pour vérifier ce fait, nous avons laissé pendant tout un hiver des manomètres constitués par des tubes de caoutchouc remplis der mercure et raccordés à des tubes de verre. Ce dispositif est assez sensible et permet par exemple de mettre en évidence la contraction du sol sous l'effet de la sécheresse. Or, dans aucun cas nous n'avons pu mettre en évidence la moindre élévation du niveau du mercure à la suite du gel. On peut donc affirmer qu'il ne se produit pas de compression.

Bien plus, il est tout à fait vraisemblable que le développement en hauteur du sol s'accompagne d'une contraction latérale provoquant des fissurations analogues à celles qui se manifestent en période de sécheresse. Ce mécanisme serait le corrolaire de la dessiccation provoquée par la formation des cristaux de glace.

Effets sur la structure du sol :

Le développement des cristaux de glace et les dilatations qu'ils produisent dans la masse terreuse ont pour conséquence de la disloquer. Cet effet est d'ailleurs bien connu des agriculteurs qui attribuent au gel la propriété de diviser la terre et utilisent ce phénomène pour préparer les terres argileuses.

Toutefois, l'effet produit est loin d'être général. Par exemple, dans nos lysimètres de Versailles remplis avec un sol contenant 20 % d'argile et 20 % de limon, nous avions un certain nombre de cases qui se sont trouvées accidentellement extrêmement humides au moment du gel pendant l'hiver 1954-55, d'autres se trouvaient complètement ressuyées. Les cases humides se sont recouvertes d'une couche de glace continue, alors que les autres se soulevaient et se fissuraient. Au dégel, la structure des premières était extrêtement compacte alors que celle des autres cases était excellente. Pour expliquer ce phénomène nous devons préciser que dans les sols à éléments fins, il ne se constitue pas de gros cristaux de glace, mais les mottes qui sont à la surface du terrain ont tendance à se développer en hauteur par suite de la formation de très fins cristaux dans la masse. Il y a ainsi une véritable dislocation des particules terreuses. Lorsqu'un vent sec soufflle sur ces formations, leur pourtour se dessèche très rapidement, la glace se sublimant dans l'air. Il se constitue ainsi une couche de terre fine remarquablement divisée, alors que la masse de terre elle-même fissurée par les effets de dessiccation, dont nous avons parlé, reste dans un excellent état de division. Au contraire, lorsque la terre est recouverte d'une couche de glace les particules qui peuvent se trouver disloquées sont immergées au dégel dans une masse d'eau qui provoque la formation d'une couche de boue et il se constitue ainsi une structure compacte.

D'une façon générale, les bons effets du gel ne s'observent que si il agit sur une terre motteuse dont les éléments sont par conséquent en mauvais contact avec la réserve d'eau du sol. Il se développe alors des cristaux lamellaires, la terre n'est jamais soumise à un excès d'eau et reste divisée au moment du dégel. Ceci est encore plus vrai si le phénomène se produit par temps sec. Nous avons pu montrer cette action de l'eau en soumettant au gel des fragments de terre malaxée, disposés soit

au contact du sol, soit sur des feuilles de papier. En les prélevant alors qu'ils sont gelés et en les immergeant dans l'alcool on peut alors dissoudre les couches de glace et recueillir les agrégats séparés qui sont relativement peu altérés par l'alcool. On constate que, pour la terre gelée, sur le papier on obtient un pourcentage nettement plus important de fragments grossiers montrant ainsi que la division est moins poussée. C'est cette fragmentation grossière qui parait être la plus avantageuse pour l'agriculture.

Conclusions :

Comme tous les phénomènes naturels, le gel du sol présente des aspects très complexes. On peut toutefois dire que ceux-ci dépendent de trois groupes de facteurs : la teneur en eau, l'importance de l'abaissement de température et la vitesse avec laquelle le gel s'établit, l'état de division et l'hétérogénéité du milieu au départ.

En milieu compact, relativement peu humide, le sol gelé est cohérent, sa structure est relativement peu divisée et l'état du sol au dégel dépendra de la teneur en eau et de la façon dont celle-ci pourra être évacuée. En milieu hétérogène, les effets de soulèvement sont beaucoup plus considérables, mais là encore l'état du sol au moment du dégel dépendra de l'humidité. Pour l'agriculteur c'est le gel sec qui paraît être le plus avantageux, surtout quand il agit sur un milieu divisé. Pour l'ingénieur, les milieux tassés, homogènes, sont certainement moins dangereux que les milieux hétérogènes, ou divisés, mais, comme l'agriculteur, il doit redouter l'excès d'eau.

Si certains aspects du gel paraissent maintenant expliqués, par contre divers mécanismes de détail, comme par exemple la formation et l'orientation des lentilles de glace, restent encore assez obscures. Il en va de même des raisons qui provoquent les mouvements des pierres à l'intérieur du terrain. Pour l'agriculteur, il reste encore à comprendre le mécanisme de la dégradation de la structure quand le gel se produit en milieu très humide. Quoi qu'il en soit, on peut dès à présent souligner que c'est l'humidité qui commande l'effet définitif produit par le gel aussi bien sur les roches que sur les sols.

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STUDIES ON LAKE-ICE MOVEMENTS

MALIN SUNDBERG-FALKENMARK
Swedish Meteorological and Hydrological Institute (SMHI)

Summary

The phenomenon that lake-ice is often pushed up the shores of a lake has been explained to be an effect of the refreezing of waterfilled contraction-cracks in the ice, ensuing from the cooling of the upper layers of the ice-sheet at low air temperature. The movements observed by Hamberg[2] in 1918, prove to be as much as 50% of the theoretical maximum value.

The refreezing of cracks of other origin, too, ought to cause similar ice movements. Investigations in a lake in Middle Sweden, where the daily variations in the water-level give rise to cracks along the shore, favour this theory. In the lake in question, however, the movements at all places observed, were directed outwards from the shore. This is explained by the fact that there existed zones of weak ice in the ice-sheet.

The expansion due to the refreezing of shore cracks, appeared as movements $\perp$ shore, which were largest where the zone of weak ice was most pronounced; these movements were of about the same size year after year. On the other hand, the movements caused by the refreezing of waterfilled contraction-cracks were not very considerable, except in the winter 1953/54 when the snow cover was thin and then only at the observation places where expansion of large parts of the lake was registered. The movements $\parallel$ shore were as a rule smaller than those $\perp$ shore.

Finally, it has been possible to obtain an approximate connection between the ice-movements $\perp$ shore and the variations in the water-level. The movements caused by the refreezing of shore-cracks have thus proved to become larger if these variations are increased.

Introduction

It is often seen, that during the winter the ice on lakes gradually moves up the shore, where it will sometimes cause a transport of shore-material (Hansen 1949[1]). The phenomenon has been explained to be an effect of the expansion due to the refreezing of the contraction-cracks caused by the cooling of the upper layers of the ice-sheet. The total shoreward movement during a winter would then be equivalent to the total width of all refrozen cracks. Evidently, the temperature of the ice surface being the same both at freezing and melting, the total movement cannot be an effect of thermal expansion.

Earlier observations of ice movements caused by contraction-cracks

Observations of such ice movements were made in Sweden in 1918 by Axel Hamberg on Lake Sommen[2]. The observations were made during the period March 2nd — 17th, when the ice was almost snow-free and thus exposed to the variations in the air temperature. He found : 1) a gradual ice movement against the shore of 40 cm in all; 2) that the main movements took place in daytime. The magnitude of the movements can evidently not be due to thermal expansion only, since the air temperature raised from —1°C to +6°C during the period, whereas the temperature of the ice surface can have raised by 1°C at the most. This rise of temperature corresponds to a thermal expansion of the surface layer of the ice-sheet (width of the lake c. 500 m, coefficient of linear thermal expansion $\alpha = 5 \cdot 10^{-5}$) of only 0.025 m.

Provided the tensile stresses in the ice, set up by the cooling of the ice in the night, exceed the strength of the ice-sheet, so that contraction-cracks are formed, the maximum effect of the refreezing of these cracks after their being filled with water, can easily be computed to

$$\sum_1^n \Delta l_i = \sum_1^n l_{i-1} \cdot \alpha \cdot \Delta t_i$$

where l_0, l_1 ...l_n are the width of the lake at the place in question at the starting observation (0), at the first measurement of the movement of the ice (1), at the second one (2) etc.; $\Delta l_i = l_i - l_{i-1}$; $\Delta t_i = t_{im} - t_i$ where t_{im} is the maximum temperature on the i: th day, t_i the minimum temperature during the preceding night. Since $\Delta l_i << l_i$, so that we can put $l_i = l_{i-1} = \ldots l_1 = l_0$, the formula can be written

$$\sum_1^n \Delta l_i = l_0 \, \alpha \sum_1^n \Delta t_i .$$

The maximum expansion during the above-mentioned Hamberg-observations, can thus be computed to be 79 cm, provided the expansion at the 500 m wide part of the lake resulted in equal movements at both shores. The value observed, 40 cm, is then c. 50% of the maximum value. There are various reasons for its being lower than 100% as, for instance, that the ice will endure certain tensile stresses without cracking, or that some cracks are not filled with water.

Studies of ice movements caused by other kinds of cracks

Hamberg's studies seem to show that large effects are produced by the refreezing of contraction-cracks; it might then be expected that other kinds of frozen cracks, e.g. the cracks along the shore caused by successive variations in the water-level should make similar effects.

When the water-level falls, the ice-sheet is subjected to bending forces produced in the region between the floating ice and the ice resting on the shore, and open cracks, which as a rule refreeze if filled with water, are formed. When the water-level rises again, the ice-sheet has become «too large», so that compression stresses occur. When exceeding the strength of the ice-sheet or that of the frozen bond ice-shore, these stresses must cause certain movements. In addition to this hypothetical type of movements there must also be a type produced by the refreezing of contraction-cracks. This latter type will in the following be called « *temperature movement* » or «*temperature effect*».

The expected movements must be directed against the region where the strength is least, i.e. against zones of weak ice, or, if such zones do not exist and the strength of the bond ice-shore is first exceeded, against the shore. That was probably the case in Lake Sommen.

Observations of ice movements in Lake Liten

After the regulation of the discharge from Lake Liten had begun, there were such damages on the timber-guiding-structures in the lake in winter, that they were supposed to be connected with the diurnal variations in the water-level caused by this regulation. Observations have been made by the Swedish Meteorological and Hydrological Institute since the winter of 1953/54, in order to make clear the real cause of the damages.

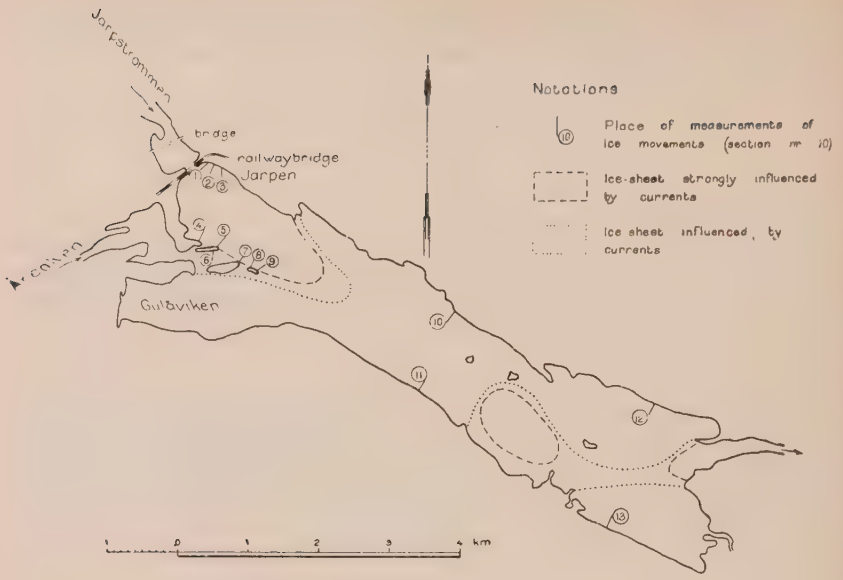


Fig. 1 — Map of Lake Liten.

Fig. 1 shows a map of Lake Liten in the Indalsälven basin in Middle Sweden. In the NW part of the lake there empty the two main tributaries, Järpströmmen and Areälven. The water from the former being fairly warm, the lake does not freeze over for c. 1 km below the railway bridge. Outside the latter tributary, however, the lake gets covered with ice. The outlet is in the SE part of the lake. The water flow through the lake is fairly large and causes pronounced currents, which impede the increase of the ice-thickness in the areas marked in fig. 1 (Melin 1948 [3]). The damaged structures are situated near to the two inlets.

The regulation of the discharge causes the following variations in the water-level: *On weekdays* there is a fall in daytime followed by a rise in the night; *on holidays* the water-level rises gradually from Saturday night till Monday morning (this is also applicable to Christmas time and other festivals when the consumption of electricity is cut down). During the week there is a gradual fall of the water-level; on this fall are superposed the above-mentioned diurnal variations. The greatest change in the water-level generally occurs on Sundays and it has not exceeded 0.35 m during the observation seasons.

In order to determine the movements of the lake-ice, measurements were made 2-4 times a month at a number of places, both in the critical upper region of the lake and in the lower parts (cf. fig. 1).

The ice-movements in each observation place (section) have been determined in relation to a system of polar coordinates, the origin of which is a stake on the shore, B, and the fix axis of which is a sight line, normal to the shore, through B and another stake higher up on the shore, A (cf. fig. 2).

The situation of the ice is represented by another couple of stakes, C and D, lying at such distance from the shore that C comes outside the zone where cracks caused by fall of waterlevel, here named *shore-cracks*, occur. D lies 30-50 m outside C. The movement of C is determined by measuring the radius vector B-C and the normal to the fix axis d_C; the movement of D by measuring the distance between C

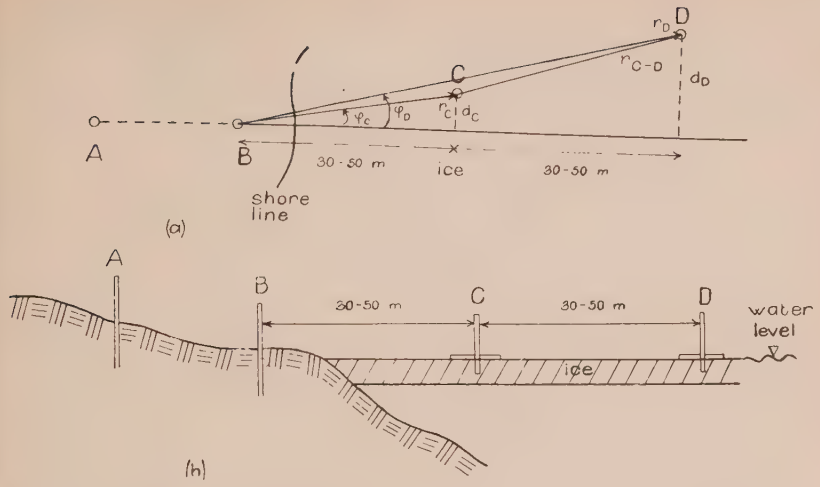


Fig. 2 — Measuring arrangements:
a) the system of polar coordinates,
b) a profile.

and D, and the normal d_D. Moreover, the level differences of the observed points have been levelled when needed for computing the horizontal distances. The measurements began as soon as the ice was thick enough to bear a person, and went on as long as possible in spring. In 1954, concentrated measurements were made daily during a ten-days' period.

The observation material has then been adapted to give the horizontal projections, $\overline{B\text{-}C}$ and $\overline{C\text{-}D}$, on the fix axis. The following corrections were made for this purpose

correction for level differences between B, C and D
» » movements normal to the fix axis (d_C, d_D)
» » thermal changes of the tapemeasure.

When computing, the observations have been assumed to have a maximum error of $\pm$ 0.01 m. Changes in $\overline{B\text{-}C}$ and $\overline{B\text{-}D}$ have been called *ice movements normal to* ($\perp$) *the shore* (written $\Delta\overline{B\text{-}C}$, $\Delta\overline{B\text{-}D}$); changes in d_C and d_D *ice movements parallell to* ($\parallel$) *the shore* (written Δd_C, Δd_D).

In table 1 are given the total ice-movements at C (represented by $\Delta\overline{B\text{-}C}$ and Δd_C) and the total ice-expansions between C and D (represented by $\Delta\overline{C\text{-}D}$) during the three winters 1953/54, 1954/55 and 1955/56. The gradual increase of that part of the movement, that was directed normal to the shore ($\Delta\overline{B\text{-}C}$), is shown in fig. 3.

Primary conclusions

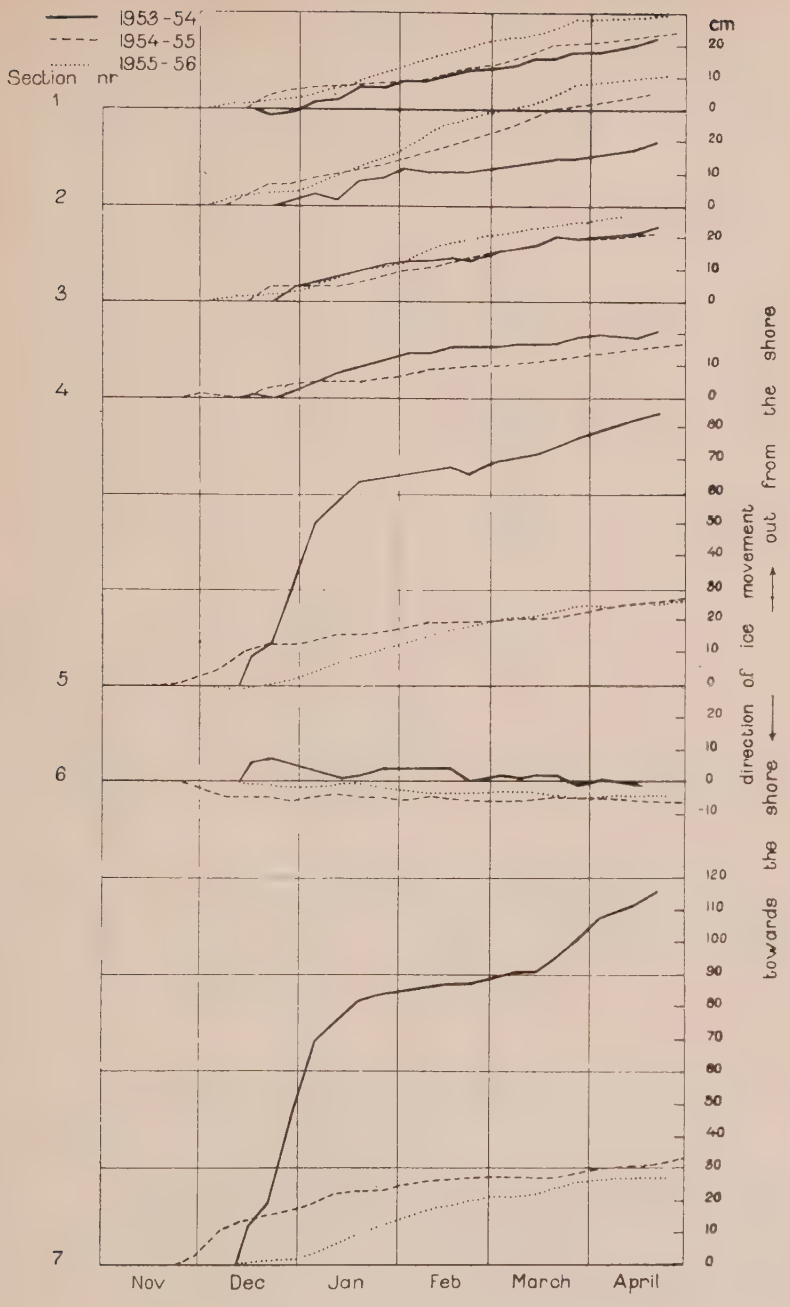
The values of table 1 show that the ice-expansion in the region B-C ($\Delta\overline{B\text{-}C}$) is considerably greater than that in the region C-D ($\Delta\overline{C\text{-}D}$). This seems, however not to be applicable to the winter 1953/54, but by comparing the lengths B-C during the different winters it will be found, that in those places where $\Delta\overline{C\text{-}D}$ was especially large in 1953/54, the length B-C was shorter than in the following years. This would

TABLE 1

Measuring lengths and total ice movements during the winters 1953/54-1955/56

1 = 1953/54 2 = 1954/55 3 = 1955/56

Measuring place (section nr)	Measuring lengths in m. B-C			C-D			Total ice movements at C $\Delta\overline{B\text{-}C}$ cm 1 against out from shore	2 against out from shore	3 against out from shore	Δd_c cm 1 with against current	2 with against current	3 with against current	Total ice expansion between C and D $\Delta\overline{C\text{-}D}$ cm 1	2	3
	1	2	3	1	2	3									
1	30	25	25	15	20	20	22	24	29	20	13	22	6	2	3
2	16	20	25	30	25	25	20	35	41	0	6	3	13	–1	2
3	30	28	30	30	45	30	24	22	27	14	13	11	4	0	–1
4	40	45	—	45	45	—	21	18	—	8	4	—	29	1	—
5	40	45	45	45	45	45	85	28	26	12	16	3	17	4	2
6	40	30	30	45	45	30	3	6	4	18	5	8	3	0	1
7	40	45	45	45	45	45	116	33	27	24	7	7	13	1	1
8	40	45	—	45	45	—	156	43	—	48	8	—	20	1	—
9	40	45	45	45	45	45	231	39	24	22	24	12	9	0	3
10		45	45		45	40	—	5	8	—	1	0	—	2	0
11			40			45	—	—	3	—	—	1	—	—	2
12			45			45	—	—	4	—	—	1	—	—	–1
13			40			40	—	—	13	—	—	1	—	—	–2


1953-54
1954-55
1955-56
Section nr
cm
1
2
3
4
5
6
7
direction of ice movement
out from the shore
towards the shore
Nov
Dec
Jan
Feb
March
April

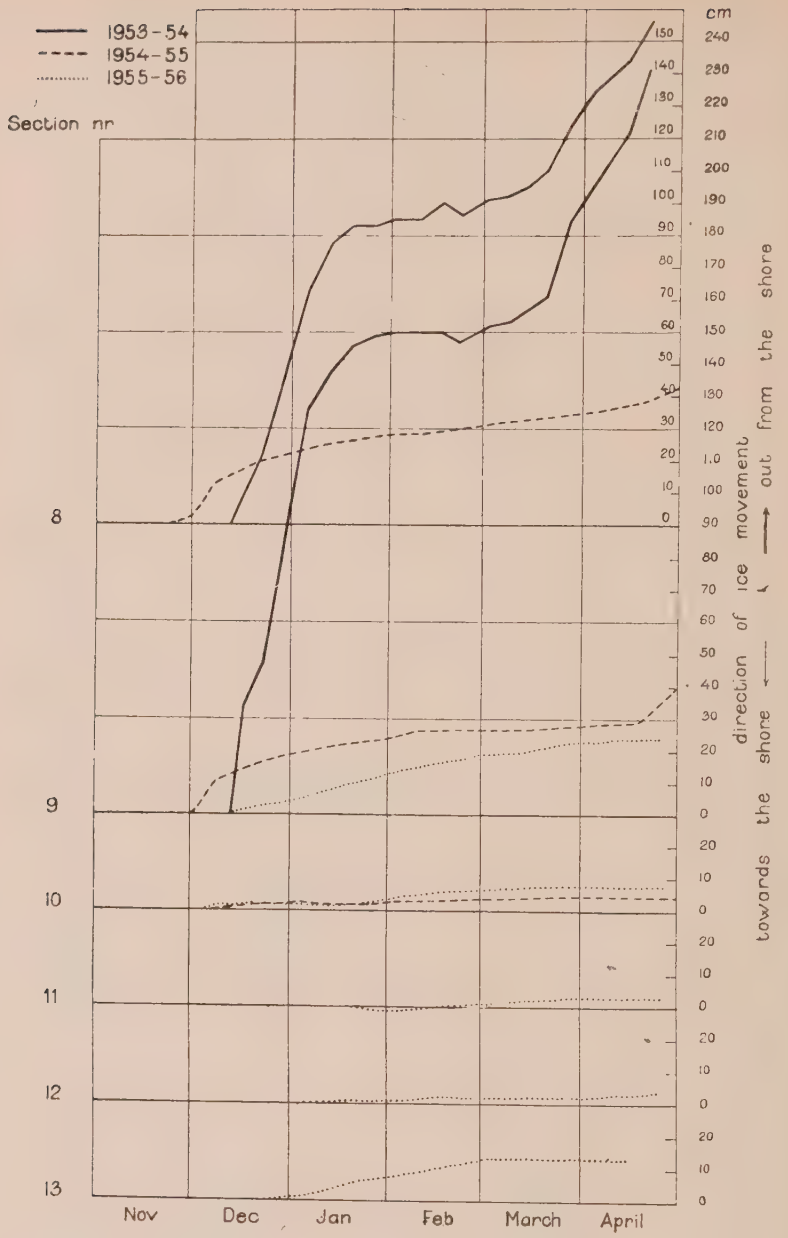


Fig. 3 — Gradual increase during the winters 1953/54-55/56 of the ice-movement ⊥ shore in the sections 1-13.

mean that the first year, C was placed so near the shore that some of the shore cracks occurred in the region outside C. The observations then show that *ice movements are generated in the region where shore cracks are formed.*

Furthermore, the observations confirm that the ice-hole at the inlet and the zone of weak-ice below, really form a decisive weak-zone, as the movements ⊥ shore were directed throughout *outwards from* the shore.

Method for determining the location of the weak-zone

The method used for controlling that the decisive weak-zone really was situated as mentioned above throughout the winter, was as follows.

The individual movements ⊥ shore, $\Delta\overline{B\text{-}C}$, were divided into three separate types, the division being based on the following factors:

1. Location of weak-zones : the analysis is accomplished for each of the assumptions a) weak-zone inside C b) weak-zone outside C.

2. Temperature change of the ice-surface since the last observation; this change is assumed to be equal to the change in air temperature, except that temperatures higher than 0°C have been put equal to 0°.

3. Effect of the refreezing of cracks.

The three types are:

type I = movements of such direction that they can be explained as effects of *thermal expansion* after a temperature rise, or *thermal contraction* after a temperature fall.

type II = movements of opposite direction to type I, that can be explained as effects of the *refreezing of* contraction-cracks or *cracks* caused by fall of water-level (into this type will come movements caused by crack-refreezing only after temperature fall; in the opposite case, both the temperature rise and the crack-refreezing cause expansion, so that both effects come into type I).

type III a) movements of such direction that they can be explained neither as an effect of thermal expansion nor of refreezing of cracks (i.e. contraction effects after a temperature rise). Some of these cases might be effects of observation errors and properly belong to type I.

b) cases that remain in type III even if the observation error ± 2 cm is regarded.

The division of the individual movements ⊥ shore being made in conformity with this principle, the probable location of the weak-zone was judged by the number of cases of type III. An example of the result of such a division of the material from the measurements of 1955/56 is given in table 2. The judged locations of the weak-zones are marked by underlinings.

Interpretation of the measurement results

The above-mentioned location of weak-zones gives the following points of view on the interpretation of the movement components.

1. *movements ⊥ shore* ($\Delta\overline{B\text{-}C}$): in the sections 1-4, 10-13, these movements can only reproduce different kinds of expansion of the ice-sheet between B and C. In this case, the magnitude of the contraction-cracks can be assumed to be of about the same magnitude as the expansion of the ice between C and D, $\Delta\overline{C\text{-}D}$.

On the other hand, the situation of the sections 5, 7-9 is such, that the movements observed in these sections might be influenced also by the effect of contraction-cracks in the part of the lake called Gulåviken, SW of the isles.

TABLE 2

Division into the types I-III *of the individual changes* $\Delta\overline{B\text{-}C}$ (*observations from* 1955/56)

Measure place (section nr)	Assumed location of weak-zone in the floating ice (outside C) type				near to the shore (inside C) type			
	I	II	IIIa	IIIb	I	II	IIIa	IIIb
1	3	7	0	0	5	0	5	4
2	2	8	0	0	4	0	6	4
3	2	8	0	0	4	0	5	4
5	6	4	0	0	3	2	5	4
6	3	2	4	1	4	4	1	0
7	5	4	0	0	4	0	5	4
9	5	4	0	0	4	0	5	4
10	5	4	1	0	5	3	4	1
11	5	3	0	0	1	3	4	0
12	3	4	1	0	2	1	5	0
13	5	3	0	0	5	0	3	3

2. *movements* || *shore* (Δd_C): these movements must essentially represent movements in the longitudinal direction of the lake, i.e. mostly temperature movements.

Exceptions are

section 1 where the nearness to the railway embankment might increase the influence of shore-cracks

section 3 where there probably is a certain influence of the shore-cracks at the protruding shore of Järpen, SE of the section.

Conclusions concerning the movements ⊥ shore ($\Delta\overline{B\text{-}C}$)

1. The movements in *sections* 1-4 have remained rather constant year after year and have increased at a fairly constant rate throughout the winter, which shows that they *cannot be only temperature effects.* Thus, they have been unaffected by variations in the thickness of the snow-cover. Furthermore, the expansion $\Delta\overline{C\text{-}D}$ has been small in comparison with the movements $\Delta\overline{B\text{-}C}$, which also indicates small temperature effects from the regions B-C and C-D. The movements are, therefore, *probably caused by shore-cracks.*

2. On the other hand, the movements in *sections* 5, 7-9 prove to be, at least partly, *composed of temperature effects* from other parts of the lake than the ice-sheet between B and C, as they vary considerably from year to year; they were largest in 1953/54, when the snow-cover was thin (cf. fig. 4). Furthermore, during 1953/54 the movements principally took place before the middle of January and from the

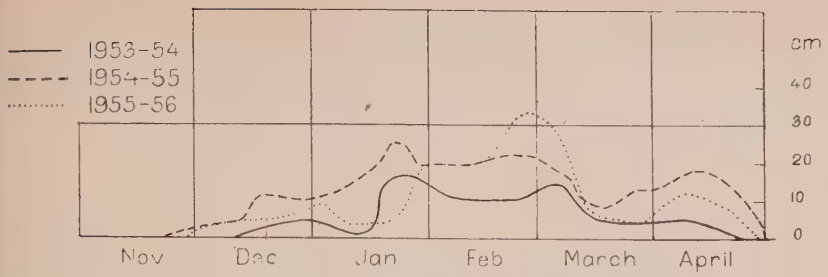


Fig. 4 — Snow depth on the ice

end of March, i.e., during the periods when the snow-cover was thinnest. Nevertheless, there were certain movements even during the period when the snow-cover was fairly thick. These movements were of about the same magnitude as the movements in sections 1-4 and could thus have been *caused by shore-cracks.*

3. The movements in *sections* 10-13 have been smaller than in the sections in the *upper parts of the lake*, but larger than the expansions $\Delta\overline{C\text{-}D}$. This might indicate, that they, too, are composed of two parts, i.e., on one hand movements due to temperature effects and on the other movements possibly caused by shore-cracks.

Section 6, see below.

Conclusions concerning the movements || shore

1. The movements in *section* 1 were fairly large and were not caused by temperature effects. They were directed down-stream as they would have been if caused by shore-cracks along the railway embankment.

2. The movements in *section* 3 were of the opposite direction and somewhat smaller than in section 1. Thus, they might have been caused by shore-cracks at the protruding shore of Järpen, SE of the section.

It is natural that the movements || shore, even if caused by shore-cracks, were less than those ⊥ shore, as the ice-sheet between the Järpen shore and the railway embankment could bear certain stresses, whereas the movements ⊥ shore, on the other hand, were directed against the open ice-hole at the inlet.

3. The movements in *section* 2 were very small, indicating that the shore-crack effects from the protruding parts of the shore cancel each other.

4. The movements in *section* 4 were also small and directed downstream, which might indicate a possible shore-crack effect from the shore protruding NW of the section.

5. The movements in *sections* 5, 7-9 varied both in magnitude and direction. During the first year they were directed downstream, but the following two years in the opposite direction.

6. The movements in *sections* 10-13 were as small as the observation error, showing that the temperature effects are small where pronounced weak-zones do not exist (cf. fig. 1).

Conclusions concerning the movements in section 6

The movements in this section prove to be *correlated with the general move-*

ments in the region SW of the ice-hole at the inlet. Thus, during the winter 1953/54, when the general movements were of a downstream direction, the movements ⊥ shore were directed outwards from the shore, whereas, during the two other winters, they were directed towards the shore. The movement || shore has been directed throughout towards the weak zone near the ice-hole at the inlet.

Special examination of shore-cracks

As has been seen above, there exist in Lake Liten ice-movements, generated by some event in the region near to the shore, and increasing at a fairly constant rate throughout the winter, obviously unaffected by the thickness of the snow-cover on the ice. In order to analyse, how this expansion develops, a ten-days' investigation with daily observations was performed in the course of Febr. 19th — 28th, 1954. A complete account of this investigation can, however, from considerations of space, not be given here, but the general outlines were as follows. In the sections 3, 5, 6, and 9 existing shore-cracks were looked for. On each side of all these cracks small stakes were hammered down into the ice. The distances between these stakes were measured daily, so that the behaviour of the cracks could be examined. Also the movements in relation to the stake B on the shore were measured. The following conclusions were made: When the water-level fell, cracks were produced near to the shore or old cracks were enlarged; when the water-level rised again, a certain amount of this expansion remained, obviously because some cracks had been filled with water and had refrozen. Some of the cracks, however, which had not refrozen, closed and, therefore, did not contribute to the expansion. The total movements ⊥ shore during a one-week-period —Sunday-Sunday—during which the total change in water-level was a rise of only 6 cm, are given in table 3; the movements have been divided into a temperature-effect-part and a shore-crack-part. The former was estimated from the changes in the regions C-D. During the observation period the ice-hole at the inlet extended down to just in front of the inlet of the Areälven.

TABLE 3

Weekly movement ⊥ shore, found by measurements during Febr. 21st — 28th 1954

Section	$\Delta\overline{B\text{-}C}$	judged temp. effect part	remaining shore-crack part
3	26 mm	11 mm	15 mm
5	33 »	19 »	14 »
6	6 »	—	—
9	24 »	19 »	5 »

These values of the weekly shore-crack effect, have been compared with the average movements per week in the ten sections in the lake. The only movements regarded are those that occurred during a standard period (the first week of December — the second week of April), when the values are comparable with respect to the influence of regulations of the discharge during holidays.

TABLE 4

Average movements per week ($\Delta\overline{B\text{-}C}$ *cm/week*)

1 = 1953/54 2 = 1954/55 3 = 1955/56

Section	Total movement $\Delta\overline{B\text{-}C}$ cm			Movement per week $\Delta\overline{B\text{-}C}$ cm/week			Code*) notation of ice state outside the section			Movement observed in Febr. 1954 cm/week
	1	2	3	1	2	3	1	2	3	
1	24	19	24	1.4	1.1	1.1	a	a	a	
2	20	28	31	1.2	1.6	1.8	a	a	a	
3	24	17	25	1.4	1.0	1.5	a	a	a	1.5
4	21	13	—	1.2	0.8	—	a	c	—	
5	72	13	23	4.2	0.8	1.3	a	c	b	1.4**)
6	–5	–1	–4	–0.3	–0.06	–0.2	d	d	d	
7	96	15	26	5.9	0.9	1.5	b	c	b	
8	134	19	—	7.9	1.1	—	b	c	—	
9	184	13	22	10.8	0.8	1.3	b	c	c	0.5
10	—	3	6	—	0.2	0.3	d	d	d	

*) a = open water outside the section throughout the winter
b = » » » » » during the greater part of the winter
c = » » » » » » part of the winter
d = icecover » » » throughout » »
**) open water down to outside this section

The table shows that the observed value 1.5 cm/week in section 3 is fairly representative; the value 1.4 cm/week in section 5, on the other hand, seems somewhat too large. This can, however, be explained by the fact that, during this observation-period, there was open water outside the section causing larger movements. The value 0.5 cm/week in section 9 seems somewhat too small compared with the value 0.8 cm/week in 1954/55, when the temperature effects ought to have been very small as the ice was covered by snow throughout the winter. In this case the judged temperature effect during the week in Febr. 1954 was probably taken too large.

In the winter of 1954/55, all the sections have lower values than in 1955/56. The difference might be a measure of the temperature effect during the latter winter, when the snow-cover was mostly somewhat thinner than in 1954-55. When comparing the magnitude of the movements during the winter of 1954/55, when there were hardly any temperature effects, it will be clearly seen, that the movements were smaller at the places where the ice was stronger and thus could bear greater stresses.

Connection between ice-movements and water-level variations in Lake Liten

The connection between the individual ice-movements in sections 1-3 and the average variation in the water-level during the three winters, has been investigated (cf. fig. 5). As the periods between the measurements are of various length, the ice-movements and the variations in the water-level have been expressed as average values per day (cm/day). The diagram indicates that a certain connection exists in spite of the rather wide distribution of the plottings; such a distribution is, however, quite natural, as:

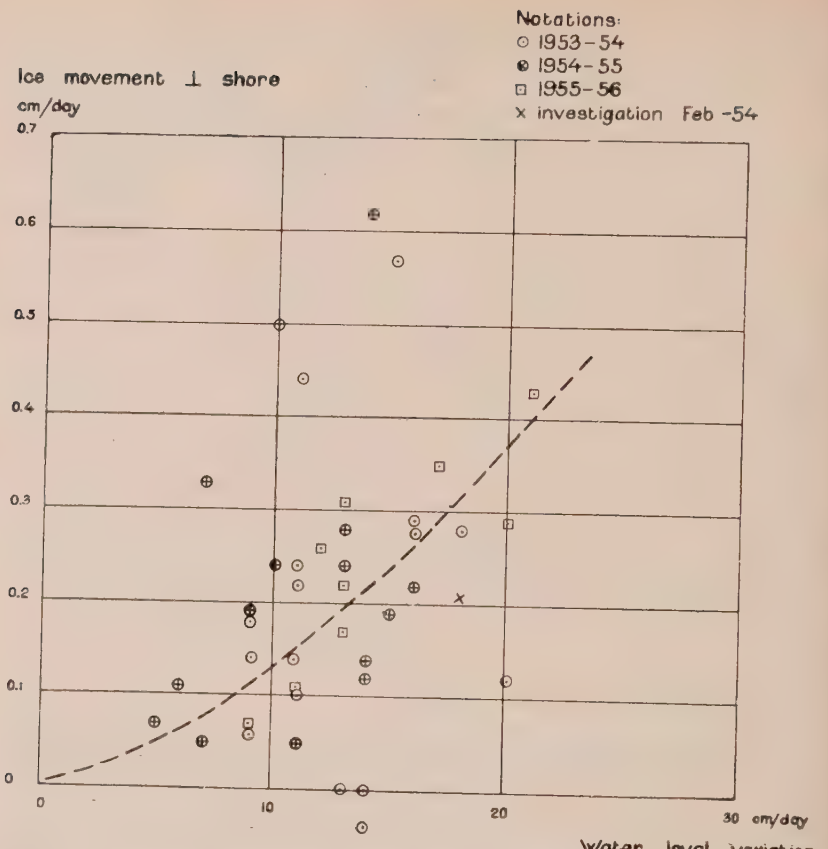


Fig. 5 — Connection between ice movements ⊥ shore and variations in the water level (sections 1-3).

1. the refreezing of the shore-cracks occurs with varying solidity, according to the weather conditions

2. the influence from temperature effects; this can be estimated from the variations in C-D. Thus, $\Delta\overline{\text{C-D}}$, in the sections 1-3 during the winters in question, did not, during a 14-days' period, exceed $\pm$ 4 cm in 1953/54, $\pm$ 1 cm in 1954/55 or $\pm$ 2 cm in 1955/56. The temperature-effect-part of the movement per day during the three winters, would then not be more than 0.28, 0.07, 0.14 cm/day resp.

A curve has been drawn to show, at least approximately, the ice-movements produced within the area in question at different variations in the water-level.

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GROUND TEMPERATURE STUDIES AT SASKATOON AND OTTAWA, CANADA

D. C. PEARCE

Abstract

An analysis is presented of the annual ground temperature data obtained at Ottawa and Saskatoon, Canada, by the Division of Building Research of the National Research Council from 1952 to 1955. The annual ground temperature variation was observed to be consistent with the theory of heat flow in a homogeneous semi-infinite solid. The observations yielded an average soil thermal diffusivity which did not vary from year to year. It was possible to include a term which approximately accounted for the insulation effect of a snow cover. Additional calculation showed the order of magnitude of the deviations to be expected from variations in the soil thermal constants.

Shortly after the formation of the Division of Building Research of the National Research Council in 1947, a study of ground temperatures was undertaken because a knowledge of the ground temperature regime was necessary to cope with several of the problems brought to the attention of the Division. The first step in this program was to establish a number of field observation stations throughout Canada. A general description of these field studies has been published ([1]). The present paper describes a detailed analysis of the observations obtained from two of the stations: one on the Montreal Road property at Ottawa, Ontario, and the other at the Prairie Regional Station at Saskatoon, Saskatchewan. These records were selected from the total available because they are reasonably complete and they extend over a period of several years.

The installation at the Montreal Road Laboratories was established specifically to study ground temperatures. Thermocouples were installed in test pits of both sand and clay to a depth of 8 feet. Certain of the test pits were cleared of snow throughout the winter to evaluate the effect of the snow cover. The details of the Ottawa ground temperature installation have been published ([2]).

The installation at the Prairie Regional Station was part of an outdoor test hut study. In order to have minimum distortion of the temperature distribution by the test huts, only the observations furthest removed from the huts were used in this analysis. At this location, 20 feet from a service tunnel, thermocouples were installed to a depth of 10 feet. No provision was made for snow clearance. Consequently, the insulating effect of the snow cover could not be independently evaluated from these data.

Description of Analysis

The ground temperature observations were studied by the mathematical techniques of Fourier analysis. The principal Fourier components of these temperature variations are the diurnal and the annual waves. However, there were insufficient data to determine the diurnal component. Consequently, this description is concerned only with the annual temperature wave.

The basic assumption was that the ground temperature at a fixed depth for

any one year could be expressed by a Fourier series of the form,

$$u(x,t) = \sum_{-\infty}^{\infty} a^n(x)\, e^{in\omega t}, \tag{1}$$

where:

$u(x,t)$ is the ground temperature at depth x and time t,
$a_n(x)$ is the amplitude of the Fourier component of frequency $n\omega$,
n is zero or an integer, and
ω is the fundamental frequency, 2π radians per year.

The functions $a_n(x)$ are given by the following integral,

$$a_n(x) = \frac{\omega}{2\pi} \int_0^{\frac{2\pi}{\omega}} u(x,v)\, e^{-in\omega v}\, dv. \tag{2}$$

These integrals were determined graphically from the observations. The precision of the data was such that reliable results could only be obtained for $n = 0, \pm 1$, which are the components of the annual wave.

This development neglects certain transient features of the ground temperature variation. This was estimated to be a small effect, however, because the amplitude of the annual temperature wave only changes by a small amount from one year to the next.

The behaviour of the annual temperature wave was then compared with the theory of steady state heat conduction in a homogeneous semi-infinite solid subject to a harmonic surface temperature. The theoretical expression for the temperature at depth x and time t is:

$$u(x,t) = A_0\, e^{-\left(\sqrt{\frac{\omega}{2k}}\right)x + i\left[\omega t - \delta - \left(\sqrt{\frac{\omega}{2k}}\right)x\right]} \tag{3}$$

where,

A_0 is the amplitude of the surface temperature wave,
δ is the initial phase angle of the surface temperature wave,
ω is the frequency of the temperature wave, and
k is the thermal diffusivity of the ground.

Equation (3) predicts an increasing phase lag and a decreasing amplitude with depth for any temperature wave. Furthermore, the magnitude of both the attenuation and the phase shift is directly proportional to the square root of the quotient of the frequency by the thermal diffusivity. Plots of the phase angle and the logarithm of the amplitude against depth for the field observations yield a check on the applicability of the theory. If this theory gives an adequate description of the behaviour of the annual temperature wave, such curves should be linear in x with a slope of magnitude $\sqrt{\omega/2k}$.

Results of the Analysis

The results indicate that the annual temperature variations at the test sites are consistent with a homogeneous semi-infinite solid model for the soil. Figures 1 and 2 are examples of phase and amplitude plots, illustrating the linear behaviour with approximately equal slopes. It also was observed that the thermal diffusivity of each

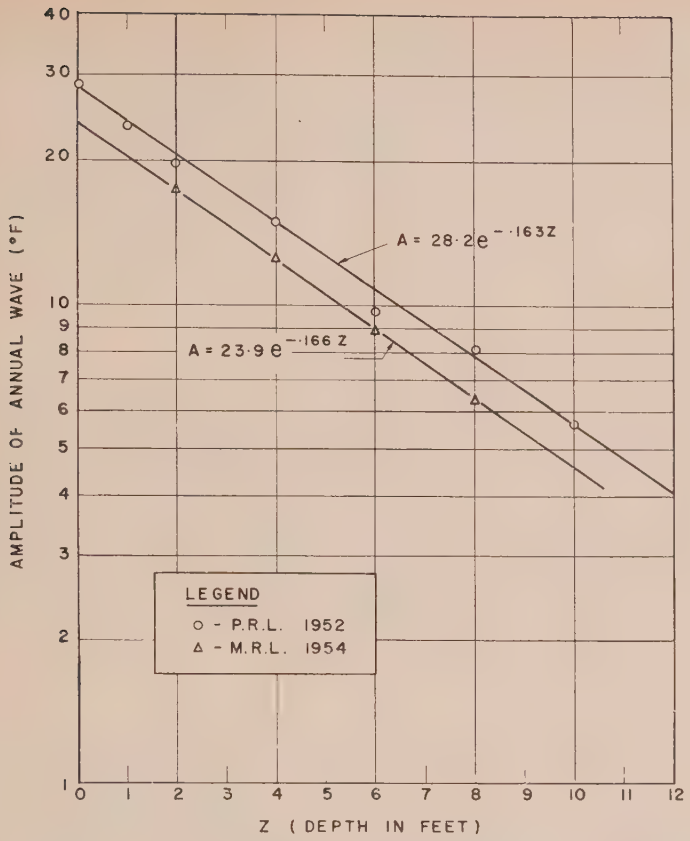

Fig. 1 — Annual wave amplitude as a function of depth.

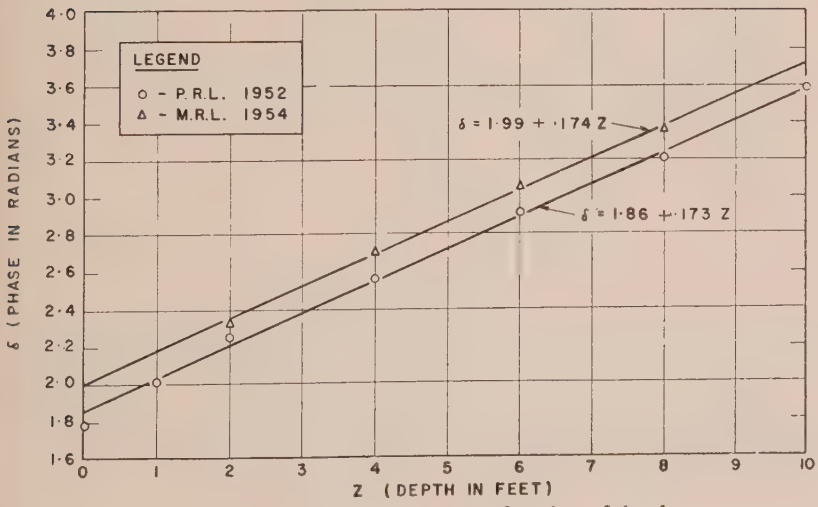

Fig. 2 — Phase of annual wave as a function of depth.

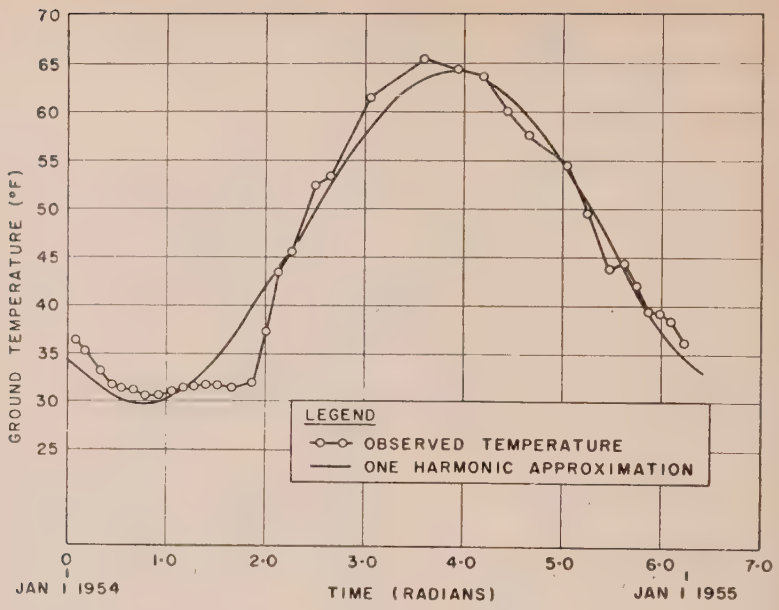

Fig. 3 — One harmonic ground temperature approximation (M. R. L. — 1954 Depth — 2 ft).

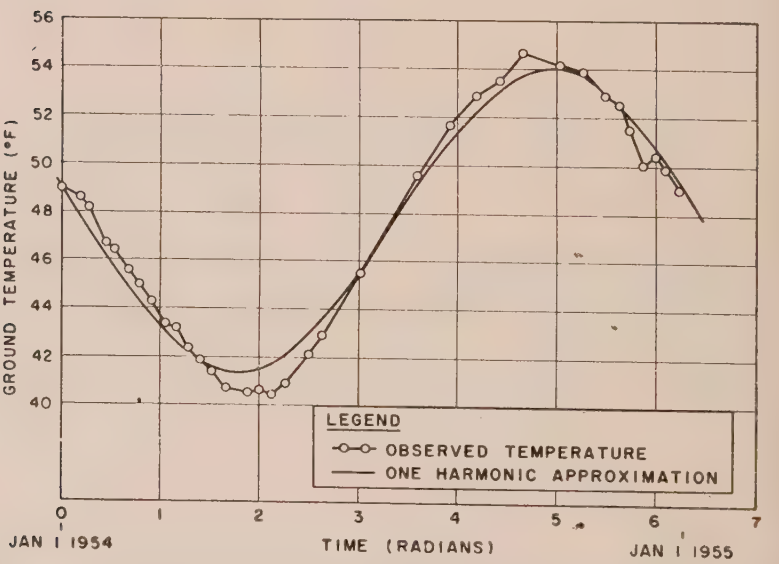

Fig. 4 — One harmonic ground temperature approximation (M. R. L. — 1954 Depth — 8 ft).

soil type considered did not vary appreciably from year to year. The value of the constant $\sqrt{\omega/2k}$ ranged from 0.13 ft.$^{-1}$ for sand to 0.17 ft.$^{-1}$ for clay. The annual wave is a good approximation to the actual ground temperature. Figures 3 and 4 give one comparison of this first harmonic approximation with the observed temperature at two different depths for the same test pit. This illustrates that the approximation improves with increasing depth. This is to be expected since the higher harmonics of the surface temperature are attenuated more rapidly with depth.

This first harmonic approximation yields a simple formula for the maximum penetration of any isotherm. The equation is:

$$x_m = \sqrt{\frac{2k}{\omega}} \log_e \left(\frac{A_0}{u_0 - u_i} \right), \tag{4}$$

where,

x_m is the maximum penetration of the isotherm u_i, and

u_0 is the mean ground temperature.

If the depth of freezing is defined as the maximum penetration of the 32° isotherm, this depth can be determined by setting $u_i = 32°$ in equation (4). The observations indicate that u_0 is approximately independent of x so equation (4) is readily soluble. Graphical solutions can be obtained if u_0 is a function of the depth.

The observations also showed the quantities $\sqrt{\omega/2k}$ and u_0 to be nearly constant from year to year. Thus the change in depth of penetration of any isotherm from year to year can be calculated from the following expression:

$$x_{1i} - x_{2i} = \sqrt{\frac{2k}{\omega}} \log_e \left(\frac{A_{10}}{A_{20}} \right) \tag{5}$$

where,

x_{1i} is the depth of penetration of the isotherm u_i for the year when the surface temperature amplitude is A_{10}, and

x_{2i} is the penetration of the isotherm u_i for the year when the surface temperature amplitude is A_{20}.

It would appear that a knowledge of the surface temperature and hence A_0 is required to use equation (4). This is not true because equation (3) affords a method of determining A_0 if the amplitude of the temperature wave is known at one depth. Thus a knowledge of the soil thermal diffusivity and a temperature variation at one depth is sufficient to use equation (4). Equation (5) is equally general. The minimum temperature for the jth year $u_{jm}(x_0)$ at an arbitrary depth x_0, in first harmonic approximation is:

$$u_{jm}(x_0) = u_0 - A_0 e^{-\left(\sqrt{\frac{\omega}{2k}}\right)x} \tag{6}$$

Thus the following relation is true:

$$\frac{A_{10}}{A_{20}} = \frac{u_0 - u_{1m}(x_0)}{u_0 - u_{2m}(x_0)}. \tag{7}$$

Equation (7) states that the ratio of the surface temperatures for any two years can be determined from a knowledge of the amplitude of the temperature at an arbitrary depth. Thus equation (5) states that in the first harmonic approximation the difference in penetration of any isotherm from year to year can be calculated from a knowledge of the soil thermal diffusivity and a time -temperature record at one depth.

Air temperature is a commonly used index of frost penetration. The mean daily air temperature near the field installations as furnished by meteorological records

was also analysed into its Fourier components. This analysis showed no simple relationship to exist between ground surface and air temperature. The amplitude of the annual component of the mean daily air temperature was greater than the annual amplitude of the ground surface while the mean annual air temperature was less than the mean annual ground temperature. Thus air temperature cannot be used directly in these formulae. However, an observed proportionality between the annual amplitudes of the ground surface and the mean daily air temperature is thought to be of some value. As a result of this proportionality, the ratio of the air temperature amplitudes for any two years can be used in place of the ratio of the surface temperature amplitudes to determine the change of isotherm penetration from equation (5).

The Effect of Variable Soil Thermal Properties

Additional calculations designed to estimate the influence of variations in soil thermal properties indicated a possible reason why the homogeneous semi-infinite model for the soil was so satisfactory. Several different variations of the thermal properties with depth were studied. The results for each analysis were quite similar. The conclusion was that the amplitude of a temperature oscillation is relatively insensitive to a change in soil thermal properties provided this change occurs within a distance that is small as compared to the length of the temperature wave. The annual temperature wave satisfies this condition so small variations of the soil thermal properties do not show up in this analysis. The calculation for one conductivity-depth dependence is given in Appendix A.

The Effect of Snow Cover

An attempt was made to include the effect of the snow cover in the theory of the annual variation of the ground temperature. Observations from the snow-covered sites indicated that a homogeneous semi-infinite model for the soil was adequate to describe the annual temperature variations. The only apparent effect of the snow cover was to reduce the amplitude of the surface temperature and thus to reduce the penetration of a given isotherm.

For this calculation it was assumed that the heat capacity of the snow was negligible and that transients associated with changes in snow depth could be neglected. Thus it is a very rough approximation to the actual situation. The main justification of this approach was that it resulted in fair agreement with the observations.

The basic idea of the analysis was to determine an attenuation coefficient characteristic of a particular winter snow cover. If the amplitude of the annual surface temperature wave without snow cover is A_0, then in the presence of snow the amplitude will be $\eta_1 A_0$. This defines the attenuation coefficeint η_1. It depends on both the depth and time of deposition of the snow cover. The method of evaluating η_1 is described in Appendix B.

Soil temperatures in a region having snow cover can be calculated from equations (4), (5) and (6) if A_0 is replaced by $\eta_1 A_0$. Thus $\eta_1 A_0$ can be considered as an effective surface temperature amplitude. The temperature distribution beneath non-snow-covered ground then is a special case with $\eta \equiv 1$.

The difference in depth of penetration of an isotherm with and without snow cover is given by the equation:

$$\Delta x_s = x_m - {}_s x_m = \sqrt{\frac{2\bar{k}}{\omega}} \log_e \left(\frac{1}{\pi} \cdot \frac{{}_s u_0 - u_i}{u_0 - u_i} \right) \qquad (8)$$

where,

x_m is the maximum penetration of the isotherm u_i in the absence of a snow cover,

${}_s x_m$ is the maximum penetration of the isotherm u_i with a snow cover,

η is the attenuation constant of the snow cover,

${}_s u_0$ is the mean ground temperature under a snow cover, and

u_0 is the mean ground temperature in the absence of a snow cover.

The Ottawa observations were compared with values calculated from equation (8). Table I gives some of these results.

TABLE I

u_i(°F)	u_0	${}_s u_0$	η	x_m	${}_s x_m$	x_s measured (ft.)	x_s calc. (ft.)
40	47.7	47.9	.878	7.7	6.8	0.9	0.9
35	47.7	47.9	.878	4.2	3.2	1.0	0.9
32.5	47.7	47.9	.878	2.9	1.2	1.7	0.8

The greatest deviation between the theory and the observations occurs for isotherms located near the ground surface. This is attributed to the contribution of the higher harmonics, which are largest in this region and which are neglected in this theory.

Conclusion

This analysis has shown that at certain test locations, the annual variation of the ground temperature is closely approximated by the theory of sinusoidal temperature variations in a homogeneous semi-infinite solid. As a consequence of the attenuation term, the approximation improves with increasing depth. Further analysis has demonstrated that the amplitude of the annual temperature oscillation appears to be relatively insensitive to variations in soil thermal conductivity. This fact probably accounts for the agreement of the observations with a theory containing only one parameter — the average soil thermal diffusivity. For the soils considered, the latent heat associated with freezing and thawing appeared to have little influence on the annual amplitude of the ground temperature oscillation. Consequently, it has been possible to develop a formula that yields maximum penetration of any isotherm. This includes the depth of frost penetration if the freezing plane is identified with 32° isotherm. These calculations require only a knowledge of the soil thermal diffusivity and a temperature record at one fixed depth. It also has been possible to include an approximation for the insulating effect of the snow cover in this formulation.

It has been a practice in predicting frost penetration to use air temperatures in place of surface temperatures. The relationship between these two temperatures has been observed to be complex so the interchangeable use of these temperatures is questionable. However, a proportionality between the annual amplitude of the air and surface temperatures was observed to exist. This was true with or without snow cover. Consequently, year-to-year changes in the maximum depth of penetration of an isotherm could be determined from air temperature amplitudes. A greater understanding of the mechanism of heat exchange at the ground surface is

required. however, to predict ground temperatures reliably from meteorologic variables.

The foregoing conclusions only apply to temperature distributions that vary slowly with time. An underlying hypothesis of the development was that heat flow in the ground obeys simple conduction theory. This assumption is not strictly accurate, however, due to the presence of water in the soil. It is quite possible that the average thermal diffusivity of the soil will depend on the time rate of change of temperature. Consequently the extrapolation of the results for the annual wave to other frequencies is not justified. A study of temperature distributions that vary more rapidly with time is required for a complete understanding of the ground temperature regime.

This study indicates that ground temperature variations of long period can be described by simple heat conduction theory. This implies that theory can be used to obtain a useful approximation to the solution of practical problems involving temperature fields that vary slowly with time. Furthermore, appropriate values for the thermal parameters for several locations in Canada are available from the field observations.

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APPENDIX A

Steady state heat concuction in l.c. semi-infinite solid with non-constant thermal properties

The purpose of this analysis was to estimate the influence of variations in the thermal conductivity on the amplitude of the temperature waves. The solution described here is for only one of several assumed conductivity-depth variations. The results for each variation were quite similar so this particular solution was assumed to contain the important features of the system. This solution is the simplest of those attempted.

Analysis of the field observations showed that the greatest deviations from the simple theory occurred near the ground surface. It was therefore concluded that the greatest variation in the thermal conductivity would also be near the surface. This fact was incorporated into the calculation by assuming that the conductivity was constant for depths greater than a fixed depth L. Furthermore, the specific heat of the soil was assumed to be constant. The steady state problem, subject to a sinusoidal surface temperature, can then be formulated as follows:

$$\frac{d}{dx}\left(K_0 g(x)\frac{df(x)}{dx}\right) - i\omega\rho c f(x) = 0,$$

where,

$K = K_0 g(x)$	is the soil thermal conductivity,
$f(x)e^{i\omega t}$	is the temperature at depth x and time t,
c	is the specific heat of the soil, and
ρ	is the density of the soil.

For this particular solution $g(x)$ was assumed to have the following form:

$$g(x) = \begin{cases} \alpha^2; & 0 < x < L \\ 1\ ; & L < x \end{cases}$$

where,

α^2 is a constant not equal to unity.

Introduce the transformation

$$u = (1 - x/L).$$

Thus the equations become,

$$\frac{d}{du}\left(g(u)\frac{df(u)}{du}\right) = \frac{i\omega\rho cL^2}{K_0} f(u) = \beta^2 f(u)$$

$$g(u) = \begin{cases} \alpha^2; & 1 > u > 0 \\ 1\ ; & 0 > u \end{cases} \tag{1a}$$

The solution of the system of equation (1a) is,

$$1 > u > 0; \qquad u = Ae^{\frac{\beta u}{\alpha}} + Be^{\frac{-\beta u}{\alpha}} \tag{2a}$$

$$0 > u; \qquad u = Ce^{\beta u} \tag{3a}$$

Assume the constant C to be fixed so that the constants A and B can be determined from the continuity conditions at $u = 0$. These conditions give:

$$C = A + B \tag{4a}$$

$$C = \alpha A - \alpha B \tag{5a}$$

Thus the solutions are

$$1 > u > 0; \quad u = \frac{C}{2\alpha}\left[(\alpha + 1)\, e^{\frac{\beta u}{\alpha}} + (\alpha - 1)\, e^{\frac{-\beta u}{\alpha}}\right] \tag{6a}$$

$$0 > u; \qquad u = Ce^{\beta u} \tag{7a}$$

For a constant thermal conductivity the value of the temperature amplitude at the surface is given by

$$A_1 = |Ce^{\beta}| \tag{8a}$$

For the selected variation of the conductivity, the surface amplitude is given by

$$A_2 = \left|\frac{C}{2\alpha}\left[(\alpha + 1)\, e^{\beta/\alpha} + (\alpha - 1)\, e^{-\beta/\alpha}\right]\right| \tag{9a}$$

The difference between A_1 and A_2 is a measure of the change of the temperature distribution caused by the non-constant thermal conductivity.

Select the phase so that C is a real number. Then

$$A_1 = Ce^{\theta} \ ; \ \theta = \left(\sqrt{\frac{\omega\rho c}{2K_0}}\right) L \tag{10a}$$

and

$$A_2 = \frac{(\alpha + 1)}{2\alpha} Ce^{\theta/\alpha} \ (1 + 2\gamma \cos\frac{2\theta}{\alpha} + \gamma^2)^{1/2} \tag{11a}$$

where

$$\gamma = \left(\frac{\alpha - 1}{\alpha + 1}\right) e^{-\frac{2\theta}{\alpha}}$$

Thus for a fractional change in K, given by $y(\alpha) = (\alpha^2 - 1)$, the corresponding fractional change in the surface amplitudes is given by

$$\chi(\alpha,L) = \frac{A_1 - A_2}{A_1} = 1 - \frac{(\alpha + 1)}{2\alpha} e^{\left(\frac{1}{\alpha} - 1\right)\theta} (1 + 2\gamma \cos \frac{2\theta}{\alpha} + \gamma^2)^{1/2} \qquad (12a)$$

If the values,

$\theta/L = 0.17$ ft.$^{-1}$ (observed value for clay)

and L = 2 ft.

are used, the following results are obtained

$y(\alpha)$	$x(\alpha,L)$
25%	1.4%
100%	16%

Thus, for the annual wave, significant changes in the conductivity are required to produce measurable changes in the surface amplitude. However, it should be noted that $x(\alpha,L)$ increases both with frequency and with L. In general, a larger value of the ratio of the distance L to the length of the temperature wave will result in a larger perturbation of the temperature amplitude.

Appendix B

Harmonic temperature variations in snow-covered ground

One of the assumptions on which this analysis is based is that the heat storage in the snow can be neglected compared to the heat storage in the ground. Thus the differential equation of heat flow in the snow is,

$$\frac{\partial^2 u(x)}{\partial x^2} = 0 \qquad (1b)$$

with boundary conditions

$$u(0) = u_i \; ; \; u(-l) = u_0 \qquad (2b)$$

$u(x)$ is the temperature in the snow cover
$x = 0$ is the ground snow interface
$x = -l$ is the upper surface of the snow.

The solution of equations (1b) and (2b) is

$$u(x) = (u_i - u_0)\, x/l + u_i$$

Thus heat flow through the snow is

$$G = -K_s \frac{du(x)}{dx} = K_s \frac{(u_0 - u_i)}{l}$$

where K_s is the thermal conductivity of the snow. Thus, to this approximation, the snow behaves as a film with a film coefficient $h_s = K_s/l$. This coefficient is time dependent because the snow depth «l» varies with the time.

The boundary value problem of heat flow in the ground, subject to this surface

film condition is formulated as follows:

$$\frac{\partial^2 v\,(x,\,v)}{\partial x^2} = \frac{\rho c}{K_g}\frac{\partial v}{\partial t}\ (x,t)\ 0 \leqslant x < \infty \tag{3b}$$

$$\lim_{x \to \infty} v\,(x,t) = 0 \tag{4b}$$

$$-K_g \frac{\partial x}{\partial v}(0,t) = h_s(t)\,[u_0 - v(0,t)] \tag{5b}$$

where

K_g is the thermal conductivity of the ground

$K_g/\rho c$ is the thermal diffusivity of the ground.

This is an extremely complicated system of equations due to the time dependence of the coefficient $h_s(t)$. To carry the analysis further, it was necessary to resort to crude approximations whose major justification was in the agreement that was obtained with the field observations.

The approximation consists of a complete neglect of the thermal transients associated with changes in the snow depth. Thus, a change in $h_s(t)$ will result in an instantaneous change in $v(x,t)$. Consequently h_s may be treated as a constant so the boundary value problem becomes quite simple. For the steady state periodic solution

$v(x,t) = W(x)\,e^{i\omega t}$ with a surface temperature

$v(0,t) = A_0\,e^{i\omega t}$ the equations become,

$$\frac{d^2W(x)}{dx^2} - \frac{i\omega\rho c}{K_g}W(x) = 0 \tag{6b}$$

$$-\frac{dW(0)}{dx} = \frac{h_s}{K_q}(A_0 - W(0)) \tag{7b}$$

The solution of this system of equations is,

$$v(x,t) = \frac{A_0\,e^{i\omega t - \left(\sqrt{\frac{i\omega\rho c}{K_g}}\right)x}}{\left(1 + \frac{K_g}{h_s}\sqrt{\frac{i\omega\rho c}{K_g}}\right)} \tag{8b}$$

Thus the temperature at the ground snow interface can be written

$$v(0,t) = \frac{A_0}{\sqrt{1 + 2\theta + 2\theta^2}}\,e^{i\left(\omega t - \tan^{-1}\left(\frac{\theta}{1+\theta}\right)\right)} \tag{9b}$$

$$\text{with}\quad \theta = \frac{K_g}{h_s}\sqrt{\frac{\omega\rho c}{2K_g}}.$$

Finally, the phase shift resulting from the snow was neglected. Thus, to this approximation the ground surface temperature is given by the equation,

$$v(0,t) = \frac{A_0}{\sqrt{1 + 2\theta + 2\theta^2}}\,e^{i\omega t} \tag{10b}$$

Equation (10b) can be rewritten,

$$v(0,t) = \zeta(t)\,A_0\,e^{i\omega t} \tag{11b}$$

where $\zeta(t)$ is an attenuation function for the snow cover given by the relation

$$\zeta(t) = (1 + 2\theta + 2\theta^2)^{-1/2}.$$

$\zeta(t)$ can be calculated from a knowledge of the snow depth and the thermal properties of the soil and snow.

Equation (11b) gives an effective surface temperature to be used in the theory of heat flow in a semi-infinite solid. The attenuation constant, η defined previously can be calculated from $\zeta(t)$. A description of this calculation follows. Assume the annual component of the surface temperature is given by the equation

$$u(t) = A_0 \sin \omega t + b_0 \cos \omega t \tag{12b}$$

In the absence of a snow cover, $u(t)$ is what has been previously called the ground surface temperature. However, with snow cover, the ground snow interface temperature is given by,

$$u_s(t) = \zeta(t)\, a_0 \sin \omega t + \zeta(t)\, b_0 \cos \omega t \tag{13b}$$

The function described by equation (13b) is not a simple sine wave. The annual component of $u(t)$, $u_{sa}(t)$ was then determined. This annual component is then represented by the equation:

$$u_{sa}(t) = a_1 \sin \omega t + b_1 \cos \omega t \tag{14b}$$

The constants a_1 and b_1 are given by the following expressions:

$$a_1 = \frac{\omega}{\pi} \int_0^{\frac{2\pi}{\omega}} (a_0 \sin \omega t + b_0 \cos \omega t)\, \zeta(t) \sin \omega t\, dt$$

$$b_1 = \frac{\omega}{\pi} \int_0^{\frac{2\pi}{\omega}} (a_0 \sin \omega t + b_0 \cos \omega t)\, \zeta(t) \cos \omega t\, dt$$

The constants a_1 and b_1 were determined by numerical integration.

The attenuation coefficient η can be determined from equations (12b) and (14b) It is given by the expression:

$$\eta = \sqrt{\frac{a_1^2 + b_1^2}{a_0^2 + b_0^2}} \tag{15b}$$

The values of η presented in Table I were determined by such a calculation. The value of the thermal conductivity of the snow was obtained from the work of Yosida et al*. A snow density of 0.28 gm/cm^3 was assumed to be an average value for settled snow. The thermal diffusivity of the soil was obtained from the ground temperature analysis. The heat capacity of the soil was calculated by the method suggested by Kersten** using field measurements of the soil density and the soil water content. Finally the soil thermal conductivity was calculated from the heat capacity and thermal diffusivity.

* Yosida, Z. and H. Iwai. Measurement of the thermal conductivity of snow cover. Snow, Ice and Permafrost Research Establishment, *Translation* No. 30. Nov. 1954.

** Kersten, M. S. Thermal properties of soils. Highway Research Board, Special Report No. 2, 1952. p. 161.

LES GLACIERS DU DESERT CHILIEN

L. LLIBOUTRY (*), O. GONZALEZ et J. SIMKEN
(*Centro de Glaciólogia de la Universidad de Chile*)

Résumé

Les alpinistes chiliens ont découvert au cours des dernières années quelques petits glaciers sur les plus hauts sommets du nord du Chili, malgré l'aridité extrême de la région. En particulier au Llullaillaco (24°43′ lat. S, 6723 m) existe un champ de glace du côté ouest entre 6500 et 5600 m; à l'Ojos del Salado (27°07′ lat. S, 6885 m) deux glaciers se trouvent sur le flanc nord-ouest, l'un entre 6600 et 6100 m, l'autre dans la prolongation du précédent entre 5800 et 5400 m; au Tórtolas (29°56′ lat. S, 6323 m) un glacier couvre tout le flanc sud-est entre le sommet et 5600 m; etc... La direction des vents dominants et l'exposition de ces glaciers varient d'une montagne à une autre. Les précipitations, orageuses en été, semblent être dues à des vents catabatiques ascendants.

En été, la température de l'air à 6500 m s'abaisse la nuit jusqu'à —28°; à 5300 m elle reste toujours en dessous de 0° C. Pourtant la neige fond franchement au soleil entre les pénitents, formant des filets d'eau courante et des lacs surglaciaires gelés en surface. Ces lacs se substituent à un névé saturé d'eau à sa base, sans qu'à aucun moment apparaisse un plan d'eau libre.

Ces glaciers n'ont ni moraines, ni crevasses, ni autre signe de mouvement. Ils sont dans toute leur étendue, suivant les années, zone d'alimentation ou zone d'ablation. Ils n'ont pas reculé sensiblement pendant les 20 dernières années. Ces derniers caractères les rapprochent de certaines régions de l'Antarctique.

Summary

During last few years Chilean climbers have discovered some small glaciers on the highest summits of northern Chile, in spite of the utmost aridity of this region. In particular at the Llullaillaco (24°43′ lat. S, 6723 m) there is a sheet of ice on the western side between 6500 and 5600 m; at the Ojos del Salado (27°07′ lat. S, 6885 m) two glaciers lie on the north-western slope, one between 6600 and 6100 m, the other one in the prolongation of the first between 5800 and 5400 m; at the Tórtolas (29°56′ lat. S, 6323 m) a glacier spreads all over the south-eastern side between the top and 5600 m; etc. The direction of the prevailing wind and the exposure of these glaciers varies from one mountain to another. Precipitations, of a stormy character in summer, seem to be originated by up-slope katabatic winds.

In summer, the air temperature at 6500 m sinks during the night to —28° C; at 5300 m it is never above 0° C. Nevertheless snow melts readily with the sun between the penitents. It forms small streams of water, as well as overglacial lakes with frozen surfaces. These lakes take the place of a water-soaked névé, without formation at any time of a free water-surface.

These glaciers have neither moraines, nor crevasses, nor any sign of movement. They are in their whole area, according to the year, either accumulators or ablators. They have not receeded sensibly during the last 20 years. To this extent they resemble, somewhat, in their behaviour some regions of the Antarctica.

Expéditions dans les Andes du nord du Chili.

Nous rapporterons ici les observations faites par des alpinistes chiliens sur les hautes montagnes du Nord du Chili au cours des dernières années, en particulier par deux d'entre nous (O.G. et J.S.). Ceux qui connaissent la torpeur et l'épuisement

(*) Actuellement : Laboratoire de Glaciologie, Faculté des Sciences de Grenoble (France).

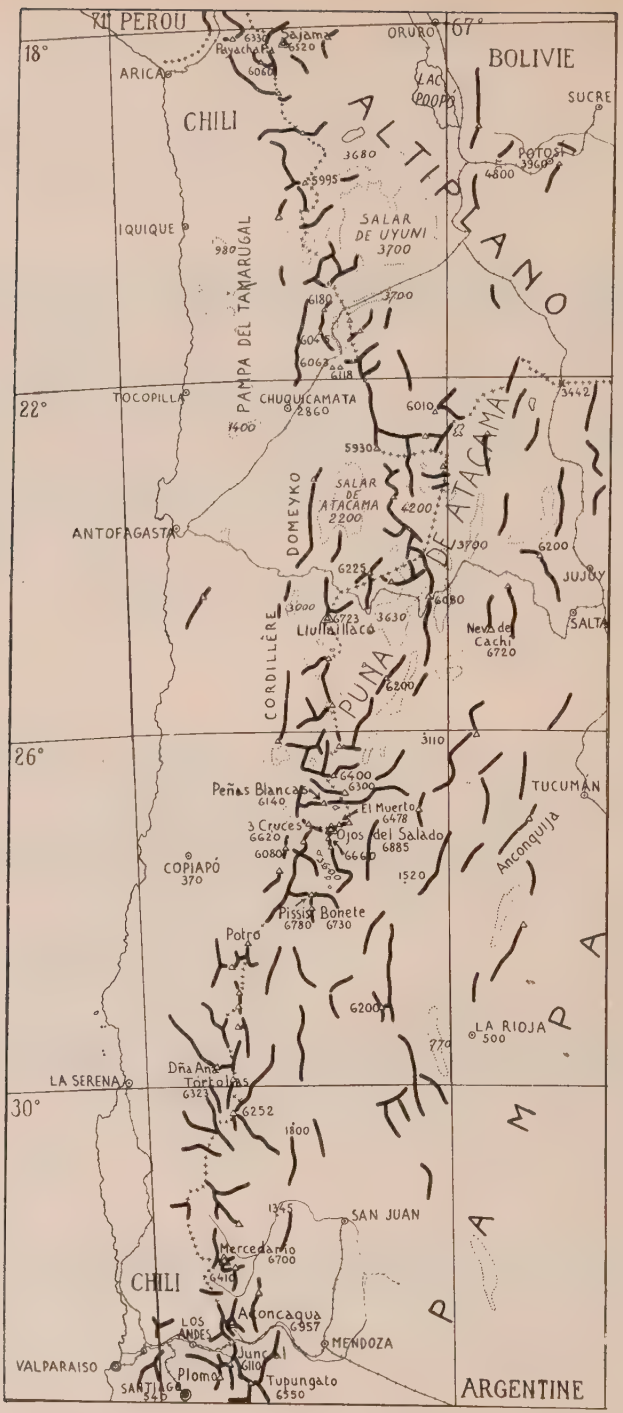

Fig. 1 — Carte des Andes du nord du Chili. De Santiago à la frontière péruvienne : environ 1700 km.

qui envahissent l'alpiniste rapidement monté à 6000 m excuseront ce que ces observations ont forcément d'incomplet. Le dépouillement des résultats et leur interprétation ont été effectués par le troisième (L.L.), après son retour en France.

Enumérons d'abord ces différentes expéditions (cf. carte fig. 1) :

Volcan Llullaillaco, 24°43 lat. S, à la frontière chiléno-argentine, 6723 m. Première ascension par Bion Gonzalez et Juan Harseim, tous deux du Club Andino de Chile, le 1er. Décembre 1952 ([6]).

Nevado Ojos del Salado, 27°07′ lat. S, à la frontière chiléno-argentine. La cote de cette montagne ayant été l'objet de controverses, précisions que les seules triangulations effectuées jusqu'à présent ont donné:

Avant 1906 (Commission chilienne des Limites) : 6880 m.

En 1937 (Expédition Polonaise) : 6863 m.

En 1956 (Expédition américaine Carter-Weaver) : 6885 m.

C'est le plus haut volcan du monde, et le deuxième sommet d'Amérique après l'Aconcagua (6959 m d'après la dernière triangulation argentine. J'avais indiqué 6957 m $\pm$ 4 m. Cf. ([8]).)

La première ascension de ce volcan avait été faite par le sud, le 27 Février 1937, par deux membres de l'expédition plonaise. La deuxième ascension, et première de l'aiguille rocheuse la plus haute parmi les quatre qui entourent le cratère, eut lieu par le nord et le nord-ouest le 5 Février 1956 ([10]). Elle fut accomplie sous la direction du Capitaine René Gajardo par une expédition de l'Ecole Militaire Chilienne de Haute Montagne, récemment créée à Los Andes. A cette expédition étaient invités plusieurs civils, entre autres l'un de nous membre du Club Andeski de Valparaiso, (Juan Simken), chargé des observations glaciologiques. Il gravit les deux aiguilles sud (6740 m) et fit le tour de la montagne à mi-pente.

Nous donnons ici une carte de la montagne (fig. 2), fragment de la carte aérienne trimetrogon au 1/250.000, agrandi, complété et corrigé.

Nevado El Muerto, 6478 m, à l'est de l'Ojos del Salado, toujours sur la frontière. Première ascension (par erreur d'ailleurs, le sommet visé étant le précédent) par Luis Alvarado, Carlos et Oscar Alvarez, et Jorge Belastino, le 23 Février 1950.

Cerro Doña Ana, 29°45′ lat. S, 20 km à l'ouest de la frontière, 5690 m. Gravi le 27 Janvier 1956 par Oscar Gonzalez et Emilio Vicens.

Cerro de las Tórtolas, 29°56′ lat .S, sur la frontière, 6323 m.
Gravi le 19 Janvier 1956 par Bion Gonzalez et Oscar Gonzalez. Comme le précédent, ce fut au cours d'une expédition du Club Andino de Chile, en collaboration avec le Musée Ethnologique de La Serena, à la recherche d'autels incaïques sur les sommets. L'un de nous (Oscar Gonzalez) y était chargé des observations glaciologiques.

Au cours de ces ascensions, au milieu de régions absolument arides, de petits névés et glaciers ont été découverts. S'ils n'ont d'autre intérêt pratique que de ravitailler en eau les alpinistes, leur intérêt scientifique par contre est très grand, car ils n'ont presque aucun point commun avec l'idée que nous avons d'un glacier de montagne. Pour simplifier le langage nous appellerons *glacier* tout champ de glace permanent (du moins à l'échelle d'une vie humaine), quelles que soient ses caractéristisques, sa forme et son mouvement; nous justifierons plus loin cette dénomination. Nous appellerons *névé* toute neige compacte, à grains arrondis,et de densité supérieure à 0,54 (porosité inférieure à 40 %), même si elle n'a que quelques mois d'âge, comme c'est vraisemblablement le cas ici.

Origine des précipitations.

Si les côtes du Pérou et du Chili septentrional sont des déserts absolus, ce n'est pas seulement parce qu'elles sont bordées par le courant froid de Humboldt, mais parce qu'elles sont à l'abri des grandes perturbations cylconiques.

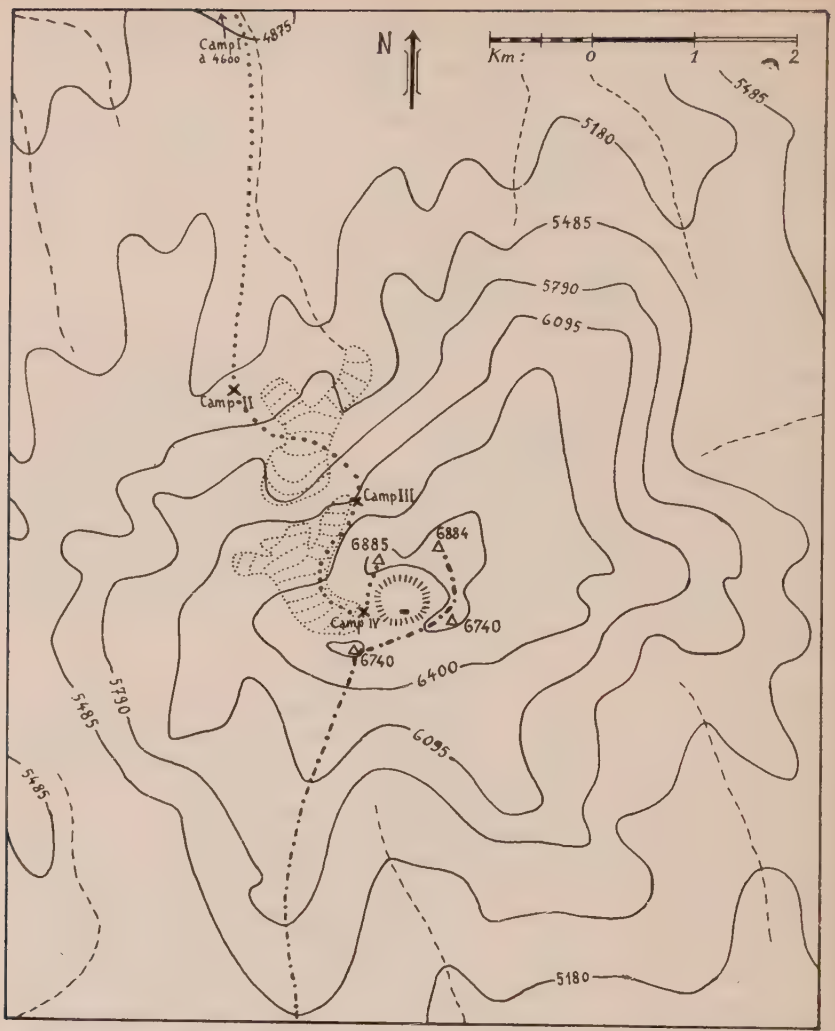

Fig. 2 — Le Nevado Ojos del Salado, avec ses deux glaciers, ses quatre aiguilles entourant le cratère (le rectangle noir indique une fumerole), et les routes d'ascension des expéditions polonaise (1937) et chilienne (1956).

De Décembre à Mars, l'air tropical humide venant de l'est amène de fortes précipitations sur la Cordillère Orientale de Bolivie, mais elles n'attaignent guère la Cordillère Occidentale Seuls, à notre connaissance, le Volcan Sajama, sommet culminant de Bolivie (6520 m) est coiffé d'une grande calotte glaciaire (2), et non loin de là, à la frontière chiléno-bolivienne, dans les Nevados de Payachata, le Volcan Pomerape (6240 m, 18°08′ lat. S) possède deux glaciers, l'un vers l'ouest et l'autre plus petit vers le sud (1).

Plus au sud, le haut plateau bolivien se prolonge dans la Puna de Atacama, avec la même altitude moyenne de 3700 m, Les nombreuses chaines de montagnes

qui la bordent à l'est reçoivent encore une certaine influence atlantique : non loin de Tucumán la Sierra de Anconquija qui culmine à 5160 m, effleure la ligne d'équilibre des glaciers actuelle, et présente plusieurs cirques glaciaires récents ([12]). Mais les nombreux sommets qui s'élèvent de plus de 2000 m au dessus de la Puna ne reçoivent en général pas assez de précipitations pour qu'un glacier quelconque puisse s'établir : il faut qu'ils la dominent de près de 3000 m. La Puna de Atacama semble former un système isolé où les vents, bien que très violents (les Polonais ont enregistré 110 km/h à leur camp de base de Tres Quebradas), sont provoqués par la topographie locale : ce sont des catabatiques.

En Février, à midi, au sommet du Muerto (6178 m), la température de l'air était —10° C. A Tres Quebradas (3800 m), le même mois, les Polonais ont enrégistré des températures de l'air de jusqu'à 21.4° C. La différence d'altitude entre les deux points est de 2620 mètres dynamiques. On voit que le gradient de température peut devenir supérieur au gradient adiabatique de l'air sec (1° C/100 m. dyn.) sur d'énormes dénivelés. L'appel d'air chaud vers les sommets explique les orages locaux et les précipitations liées à la topographie.

Sur ces volcans coniques les influences structurales sur l'emplacement des glaciers sont faibles; les radiations solaires sont proches de la verticale à midi; enfin la dissymétrie est-ouest dans l'ablation qui se produit lorsque celle-ci est due à la fois à la radiation solaire et à la chaleur de l'air ([8]) ne joue pas, l'air ne s'élevant jamais au dessus de 0°. L'on peut admettre alors que les glaciers se forment sous le vent des sommets, et déduire de l'emplacement des glaciers la direction des vents dominants.

Au camp de base de Barrancas Blancas (4600 m), au nord du Nevado Ojos del Salado, les vents dominants étaient du nord-ouest. Au nord de ce lieu s'étend une crête est-ouest, d'altitude moyenne 5500 m, dont l'un des sommets culminants (Volcan Peñas Blancas, 6020 m) a été gravi. Cette crête a huit « champs de glace » du côté sous le vent, c'est-à-dire sur le versant sud, dans de légères dépressions qu'on ne peut encore appeler cirques. Mais au sommet de l'Ojos del Salado, le vent soufflait du sud et les glaciers se trouvent sur le versant nord-ouest. Plus à l'est, au Nevado El Muerto, comme plus à l'ouest, au Nevado Tres Cruces, les glaciers sont de nouveau sur le versant sud. La région de l'Ojos del Salado semble donc bien être un point de convergence de vents opposés.

Plus au sud, dans le « Norte Chico », les vents dominants du sud-ouest qui règnent dans le Chili Central, d'origine cyclonique, font leur apparition. Mais à la différence du Chili Central et de la Patagonie où les orages électriques sont presque inconnus, en été, dans les montagnes, il y a presque tous les après-midi, entre 15 h 30 et 18 h, des chutes de neige accompagnées d'éclairs. Au Doña Ana il y a de minuscules glaciers (des « planchones de hielo », c'est-à-dire des « champs de glace») au dessus de 4800 m, et non de 4150 m comme l'affirme Von Klebelsberg pour les sommets dominant la vallée de l'Elqui. Ils se trouvent de tous les côtés de la montagne sauf au nord. Au Tórtolas il existe un glacier sur le flanc sud-est, qui descend jusqu'à 5600 m.

Vers 32° lat. S, entre le Mercedario (6700 m) te le Côrdon de la Ramada (6310 m) se trouvent les premiers grands glaciers, atteignant 12 km de long et descendant jusqu'à 4000 m ([10]). Nous quittons la zone aride tropicale pour entrer dans une zone subtropicale de climat à peu près méditerranéen (plus exactement californien), avec abondantes précipitations d'hiver, d'origine cyclonique. L'influence de la latitude et l'effet est-ouest signalé plus haut font que l'englacement y soit plus fort sur les flancs sud-est des montagnes ([5,8]).

Etat de la neige en été.

La radiation solaire intense et le refroidissement nocturne tout aussi intense par suite d'un air très sec transforment rapidement la neige en névé cohérent et y sculptent des pénitents, suivant un mécanisme que l'un de nous a décrit par aillleurs [1,8]. Les pénitents sont omniprésents sur les champs de glace ou de névé.

Une seule exception signalée : au Llullaillaco, entre 5600 et 6500 m du côté ouest, existait en Décembre 1952 un champ de glace lisse, sans névé ni pénitents. Nous l'attribuons au fait qu'il s'agissait d'une croupe de pente forte, bien exposée aux vents ascendants. La pente forte et l'exposition à l'ouest diminuent l'ensoleillement (n'oublions pas que le ciel se couvre en général les après-midi), le vent ascendant est humide, aussi l'évaporation est-elle faible, et le mécanisme de formation des pénitents ne se déclenche pas. Les mêmes faits s'observent dans les Andes de Santiago.

Au Doña Ana, en Janvier 1956, on trouvait sur les glaciers des pénitents de névé de 40 à 60 cm de haut, partiellement recouverts de neige fraîche. Au Tórtolas, mis à part un peu de neige fraîche, le glacier était à nu à cette époque, avec des pénitents de glace très étroits de 20 cm de haut. Au Nevado El Muerto, vers 6000 m, l'expédition Belastino trouva dans une légère dépression à peu près horizontale, sur plusieurs hectares, des pénitents de névé gigantesques, de 5 à 8 m de haut. Toute cette neige avait du s'accumuler l'hiver précédent, car on n'observait qu'une strate de poussières à mi-hauteur, qui était parfaitement régulière et s'était donc formée sur un champ de neige lisse. Sur du névé de plusieurs années (jamais observé d'ailleurs dans les Andes à pénitents), plusieurs strates de poussières, pertubées par les pénitents, auraient été visibles. Cette hauteur de neige accumulée en un hiver et atteignant localement 8 mètres laisse rêveur, quand on pense qu'à Copiapó il ne tombe que 2 cm d'eau par an !

En Février 1956, à l'Ojos del Salado, des pénitents de névé de jusqu'à 2 m de haut abondaient aux camps II (5300 m) et III (6100 m). Ils fournissaient des filets d'eau courante entre 10 h et 16 h. Au camp IV (6565 m) les pénitents étaient recouverts de neige fraîche en franche fusion et imbibée d'eau, « pourrie ». Pourtant les températures de l'air à ces altitudes sont toujours négatives. Par recoupement et interpolation des données très incomplètes des divers alpinistes (toujours plus soucieux de noter les températures minima que les températures maxima !), on aboutit à une température de l'air en plein été oscillant entre 0° et —17° à 5300 m, entre —10° et —28° à 6500 m. Mais dans les couloirs entre les pénitents, où la radiation solaire se concentre et où l'air se sature d'humidité, règne un microclimat tout-à-fait différent.

Les lacs gelés.

Cette eau de fonte qui se forme dans les couloirs des pénitents et qui sature la neige forme parfois de petits lacs surglaciaires perpétuellement gelés.

L'expédition polonaise à l'Ojos del Salado avait établi son camp d'altitude, dit-elle, au bord d'un tel lac gelé, vers 6400 m (probablement près du sommet innominé 6660 m au sud de l'Ojos del Salado). L'expédition Gajardo a trouvé un autre lac gelé en haut du glacier inférieur, vers 5800 m. Au Tórtolas, à la selle sud-est entre le sommet et une antécime, vers 5900 m, existe aussi un lac gelé. Au Doña Ana un lac circulaire de 10 m de diamètre occupe le sommet; il est probablement logé dans un ancien cratère volcanique.

Au Cerro Plomo (5430 m, 33°15' lat. S, 40 km au nord-est de Santiago), l'un de nous (O. Gonzalez) a observé la formation d'un petit lac gelé près du sommet, au cours de la semaine passée là-haut avec Luis Krahl en fouilles archéologiques

(c'est sur ce sommet qu'on a découvert le corps parfaitement conservé par le froid d'un petit enfant inca). Ce lac gelé s'était substitué à un champ de pénitents sans qu'à aucun moment se forme une surface d'eau libre.

Le mécanisme semble être le suivant. Dans une dépression l'eau de percolation ne peut s'écouler, forme une nappe phréatique. L'ablation ne peut aller au-delà de cette nappe, constamment alimentée par les nappes supérieures. Lorsque les pénitents atteignent cette zone, ils diminuent et disparaissent. Simultanément le dégel aux heures chaudes de la journée ne peut plus se produire, et la surface de la nappe phréatique reste gelée. La fonte cependant se produit sous une couche de glace superficielle, sous l'effet des radiations solaires qui la traversent. Un *effet de serre* (fonte au sein d'une atmosphère froide sous une couche transparente) se substitue à l'*effet de pénitent* (fonte au sein d'une atmosphère froide dans des creux). De petits micropénitents, de quelques centimètres de haut, peuvent apparaitre dans la pellicule de glace du lac gelé, mais ils ne peuvent traverser la pellicule de part en part, l'eau gelant dès qu'elle sourd à l'air libre.

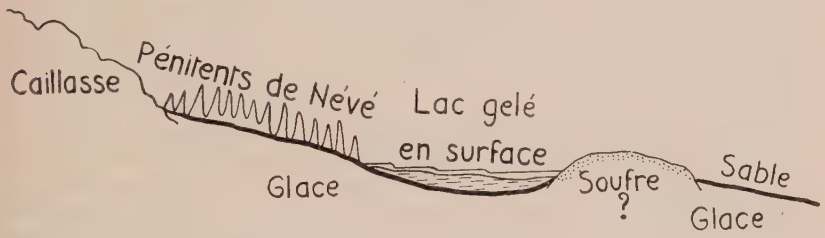

Fig. 3 — Profil schématique du lac surglaciaire gelé à 5800 m, à l'Ojos del Salado

Le lac gelé étudié par l'un d'entre nous vers 5800 m à l'Ojos del Salado, lac dont nous donnons un profil schématique fig. 3, avait environ 20 m de long, 15 m de large et 1/2 m de profondeur. Il était couvert d'une couche de glace de 5 à 10 cm d'épaisseur, avec des micropénitents sur les bords. Cette surface n'était pas toute de niveau, mais formée d'une série de terrasses horizontales en escalier, avec une dénivellation d'un décimètre entre deux terrasses successives. L'existence de ces terrasses est évidemment incompatible avec un plan d'eau libre à un moment donné.

L'eau alimentant le lac provenait des pénitents de névé des pentes supérieures (dessinés de profil sur le schéma, ce qui est inexact). Le lac devait son existence à un barrage de 1 m de hauteur, 8 m d'épaisseur et 15 m de longueur constitué par du soufre natif. L'apparition de cette butte de soufre au milieu du glacier inférieur reste un mystère. Peut-être le lac occupait-il l'emplacement d'un ancien cratère adventice du volcan, peut-être n'y a-t-il eu là autrefois qu'une solfatare?

Ces lacs gelés des hauts sommets andins font penser aux lacs gelés observés dans une chaine de montagnes de l'Antarctique, dans une région où la température de l'air est toujours négative, par l'expédition allemande Schwabenland ([3]).

Les glaciers.

Le seul glacier mentionné par von Klebelsberg entre le Sajama et le Mercedario se trouverait au Cerro Potro (5830 m, 28°24′ lat. S, à la frontière). Rappelons tous les nouveaux glaciers découverts :

Llullaillaco 6723 m, 24°23 lat. S: 1 champ de glace à l'ouest, entre 6500 et 5600.

Peñas Blancas, 6020 m, 26°49′ lat. S, 8 champs de glace vers 5500 m.

Région de l'Ojos del Salado. 27°07′ lat. S environ. D'est en ouest on trouve successivement les sommets suivants, appelés tous *Nevados*, c'est-à-dire sommets enneigés (ou englacés) :

Incahuasi, 6610 m : aucune donnée.

El Fraile, 6044 m : une tache de neige perdure au nord-ouest sous le sommet.

El Muerto, 6478 m : 2 petits glaciers presque entièrement couverts vers le sud.

Ojos del Salado, 6885 m (cf. carte fig. 2). Un glacier part à 6600 m à l'ouest du cratère, se divise en deux lobes, l'un vers le nord, l'autre vers le nord-ouest, et disparait sous les sables vers 6100 m. Dans le prolongement, entre 5800 et 5400 m se trouve un deuxième glacier de pente plus forte, qui se divise aussi en deux langues.

Sommet innominé 6660 m, en retrait des autres vers le sud : probablement non englacé.

Cerro Solo, 6190 m : une tache de neige ou de glace visible sous le sommet en 1937 avait disparu en 1956.

Tres Cruces (nombreux sommets), 6330, 6620, 6030 m, etc. : Glaciers étendus sur les versants sud et quelques glaciers de cirque sur les versants est. En comparant des photographies de ces sommets prises depuis l'Ojos del Salado en 1937 et 1956, on ne remarque aucune différence appréciable d'englacement entre ces deux dates. Nous verrons toutefois qu'il faut a priori davantage s'attendre à des variations d'épaisseur qu'à des variations d'aire.

Doña Ana, 5690 m, 29°45′ lat. S: nombreux champs de glace au-dessus de 4800 m de tous les côtés, sauf au nord.

Tórtolas, 6323 m, 29°56′ Lat. S: un glacier au sud-est entre 6320 et 5600 m.

Les glaciers signalés sont des champs de glace compacte, plus ou moins ondulés, recouverts par endroits de moraine d'ablation. Dans le glacier inférieur de l'Ojos del Salado, l'un de nous a pu observer la stratification : il existe des lits de sable tous les 30 à 50 cm, qui se cumulent vers le bas pour former une moraine d'ablation de sable. Ce sable a sûrement été amené par les tourbillons de vent. 30 à 50 cm, c'est l'ordre de grandeur de la glace de regel formée chaque printemps dans les Andes de Santiago vers 4300 m, et qui peut se conserver enterrée sous les éboulis ([7]).

L'on n'observe ni crevasse, ni autre trace de mouvement. Les glaciers sont souvent logés dans des cirques peu prononcés, entaillant légèrement la pente, de grandes niches de nivation si l'on préfère, d'autres fois ils gisent sur la pente sans cirque du tout.

Sur les cotés l'on n'observe pas de moraine, à leur sommet pas de rimaye : ils disparaissent sous les éboulis avec lesquels ils sont de niveau. Parfois l'amincissement est visible, les cailloux du fond apparaissant entre les pénitents de glace. Exceptionnellement au Tórtolas le glacier se termine à 3 m du sommet par un fossé (*Randkluft*) de 1 m de profondeur. Ce fossé est dû à la chaleur irradiée par les rochers voisins, non au mouvement.

En résumé il s'agit de champs de neige transformés sur place en glace, par regel. Quelques glaciers de ce type existent plus au sud, dans les Andes de Santiago, par exemple sur les flancs du Cerro Plomo ou du Volcan San José. Mais là ils voisinent avec des glaciers de cirque et de vallée, munis de rimaye, crevasses et moraines médianes.

Considérations générales.

Certaines lecteurs s'étonneront que nous appelions glaciers des champs de glace et de névé sans mouvement appréciable. Pour eux un glacier est un fleuve de glace par lequel les précipitations solides en excés dans les hautes régions s'écoulent vers des régions où l'ablation est plus forte. Comment appeler alors les champs

de glace observés? Car le mot « champ de glace » (*icefield*) a déja été employé pour désigner de grandes extensions de glace, de vrais petits inlandsis, se déversant par toute une série de glaciers périphériques. Et si un glacier de nos Alpes devient stagnant à la suite d'une longue série d'années déficitaires, va-t-il changer de nom en attendant des années meilleures? Il est génant de ne pas avoir un mot pour désigner toute grande masse de glace persistante dans la nature : inlandsis, shelf, icefield, icecap, glaciers de cirque, de vallée, de piémont, etc..., quelles que soient ses caractéristiques. Nous ne voyons pas d'autre terme acceptable que « glacier ».

Le schéma classique du glacier *stricto sensu*, de ce que nous appellerons un glacier évacuateur ([9]), est lié à la notion d'une *limite des neiges persistantes*, altitude à laquelle l'accumulation neigeuse devient égale à l'ablation; au dessus toutes les régions de pente faible sont englacées. Comme l'un de nous l'a montré par ailleurs ([5]) la présence des pénitents et l'intense regel de l'eau de fonte modifient déja beaucoup ce schéma. L'on ne peut plus parler que d'une *ligne d'équilibre du glacier*, où l'ablation nette sur le glacier (compte tenu du regel) égale la précipitation. Un glacier sans névé peut alors être en équilibre au milieu de pentes douces de terre nue.

Mais ce schéma modifié néglige encore les variations de la précipitation d'une année à l'autre. Dans les Andes de Santiago nous avons déja parlé de deux types d'années ([5, 8]) :

1°) Années normales : précipitations annuelles à Santiago groupées autour de 300 mm, ligne d'équilibre sur les Glaciers Olivares s'établissant vers 4600 m.

2°) Années exceptionnellement humides : précipitations annuelles à Santiago groupées autour de 660 mm, la ligne d'équilibre sur les Glaciers Olivares s'établissant vers 3800 m.

Plus au nord la dispersion des précipitations devient incomparablement plus grande. A San Juan (Argentine), la précipitation annuelle a varié, au cours des derniers 100 ans, entre 8 et 196 mm.

Le schéma classique néglige aussi les facteurs topographiques locaux. Or, comme nous l'avons dit, ce sont ces facteurs qui provoquent et déterminent les précipitations. En première approximation il faudrait parler de la précipitation pour un sommet donné, et non pour une altitude donnée.

Finalement, suivant les années et suivant le glacier considéré, le glacier sera entièrement zone d'accumulation ou entièrement zone d'ablation. S'il se maintient, ce n'est pas parce que les bilans de sa partie haute et de sa partie basse se compensent, mais parce que les bilans des années successives se compensent.

Pour étudier de tels glaciers stagnants on ne peut donc pas étudier les fluctuations de la ligne d'équilibre. Leurs variations d'aire sont en général inappréciables devant leurs variations d'épaisseur. Il faudrait choisir un point particulièrement représentatif de l'état général du glacier et y mesurer l'accumulation, l'ablation et le regel au cours du temps, c'est-à-dire faire des études stratigraphiques en un point donné. Nous doutons qu'on aille de sitôt creuser la glace à 6000 m dans la Puna de Atacama, mais ces considérations restent en grande partie valables pour l'Antarctique ou les petits glaciers des Alpes du Sud.

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Effet est-ouest dans l'ablation : p. 73.
Pénitents : p. 74-95.
Altitude de l'Aconcagua : p. 257.
Glaciers du nord du Chili : p. 305-306.
Lacs gelés à très haute altitude : p. 438.
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CARTOGRAPHIE
(*meilleures cartes à ce jour*)

Pour le relief : *World Aeronautical Chart* de l'A. A. F. au 1/250.000. Récemment traduite et rééditée par l'Institut Géographique Militaire chilien sous le nom de *Carta Preliminar* (une deuxième édition avec ombres est sous presse).

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DER EISHAUSHALT AM HINTEREISFERNER IN DEN JAHREN 1952/53 UND 1953/54

O. SCHIMPP

Summary

Measurements on 54 gauges brought dates of the seasonal fluctuations of the glacier movement. The vertical component of the movement was especially determined. The length of the tongue of the Hintereisferner does not correspond to the alimentation in the firn-area, but is the rest of a passed glacier-stand. The vertical movement, along with annual accumulation and ablation data, gives a new solution to calculate the glacier-economy separated for alimentation- and ablation-area. This shows the real relations between alimentation and economy.

Einleitung :

Die starken Veränderungen, die die Gletscher in geschichtlicher Zeit erfahren haben und besonders der auch dem Laien deutlich erkennbare Gletscherrückgang der letzten hundert Jahre, hat die Frage nach der Art und den Ursachen dieser Gletscherschwankungen aufgeworfen.

Aus den meteorologischen Daten lassen sich diese Ursachen aus mehreren Gründen nicht ablesen. Da nur an wenigen Orten seit hundert Jahren eingehende meteorologische Messungen vorgenommen werden und die Meteorologie zu diesem Zweck mit fünfzigjährigem Mittel arbeiten muß, sind die Unterlagen nicht ausreichend. Außerdem sind die Gletscherschwankungen, beziehungsweise die Änderungen der Ernährung, von einer Reihe von Faktoren abhängig, dem Niederschlag und seiner Verteilung, der Temperatur und der Strahlung. Eine kleine Änderung an einem dieser Faktoren, die durch Mittelwerte überhaupt noch nicht bestimmbar ist, kann bereits so große Schwankungen bewirken, wie sie in den letzten hundert Jahren stattgefunden haben. Da alle Faktoren gemeinsam die Gletscherschwankungen bestimmen, sind ihre Veränderungen um so kleiner und um so weniger bestimmbar. Der Umstand, daß der Gletscher bereits feinste Klimaschwankungen anzeigt, hat die glaziologische Forschung weiter an Bedeutung gewinnen lassen.

Da die Gletscherkunde heute noch nicht in der Lage ist, die Ursache der Gletscherschwankungen zu klären, ist es ihre Aufgabe, Material verschiedenster Art zusammen zu tragen, das in der Zukunft einmal eine Lösung des Problems ermöglichen soll. Die beste Möglichkeit eine umfassende und genaue Aussage über die Ernährung eines Gletschers zu machen, ist die Bestimmungs einer Massenbilanz also des Gletscherhaushaltes. Besonders H. W.: son Ahlmann hat in Wort und Schrift gefordert, Gletscheruntersuchungen so durchzuführen, daß sie eine Haushaltsbestimmung ermöglichen.

Eine neue Möglichkeit einer genauen Bestimmung der Massenbilanz ergab sich aus den Untersuchungen des Verfassers am Hintereisferner (Ötztaler Alpen) in den Jahren 1952-54. Die vertikale Komponente der Gletscherbewegung ist mit eine Arbeitsgrundlage dieser Methode. Wie unten im einzelnen ausgeführt wird hat sie den Vorteil für jeden Punkt des Gletschers, an dem gemessen wurde, gesondert die Jahresbilanz zu bestimmen. Sie erfordert aber für die Gesamtbilanz eine große Zahl genauer Messungen.

Auch die Teiluntersuchungen dieser Arbeit, besonders die der jahreszeitlichen

Schwankung der Gletscherbewegung und der vertikalen Bewegungskomponente brachten neue Gesichtspunkte. Vom Herbst 1952 bis zum Herbst 1954 wurden durchschnittlich jeden zweiten Monat 54 Pegel eingemessen, um die Gletscherbewegung in der horizontalen und der vertikalen Komponente zu bestimmen. An den selben Pegeln wurde auch die Ablation, beziehungsweise die Akkumulation gemessen. Die Pegel waren in 8 Profilen am Hauptgletscher und in je einem Profil an den drei Seitengletschern angeordnet. Die Messung erforderte sieben verschiedene Theodolitstandpunkte. Daneben konnten noch Versuche einer Messung der tageszeitlichen Änderungen der Fließgeschwindigkeit und einer Bestimmung der Bewegung am Untergrund des Gletschers gemacht werden.

Die Wahl des Hintereisferners zum Arbeitsgebiet für diese Untersuchungen erfolgte aus mehreren Gründen. An keinem anderen Gletscher der Ostalpen sind bereits seit so vielen Jahren und in solchem Umfang Forschungen durchgeführt worden. Der Hintereisferner stellt durch seine gleichmäßige Gestalt beinahe das Modell eines Gletschers dar. Die dynamischen Verhältnisse sind daher verhältnismäßig einfach. Es war außerdem eine Erleichterung der Arbeit, daß das Begehen des Hintereisferners geringe bergsteigerische Anforderungen stellt.

Im zweiten Arbeitjahr wurde diese Arbeit mit anderen Untersuchungen koordiniert und auch andere Wissenschaftler eingeladen, am Hintereisferner ihre Studien zu betreiben. Diese Arbeitsgemeingschaft erhielt eine Unterstützung durch die Bayrischen Innkraftwerke, die Deutsche Forschungsgemeinschaft und den Deutschen Alpenverein. Meine Arbeit wurde vor allem durch Prof. R. Finsterwalder mit Rat und Tat unterstützt. Der Deutsche Alpenverein stellte mir als Messinstrument einen Wild T2 zur Verfügung. Der Treuhandverwalter der deutschen Alpenvereinshütten, Hofrat M. Busch gewährte große Vergünstigungen für den Aufenthalt am Hochjochhospiz.

Die Gletscherforschung am Hintereisferner:

Durch eine große Zahl von Untersuchungen, die für die Gletscherkunde von grundlegender Bedeutung sind, ist der Hintereisferner allgemein bekannt geworden Im folgenden soll die Bedeutung dieser Arbeiten kurz unterstrichen werden.

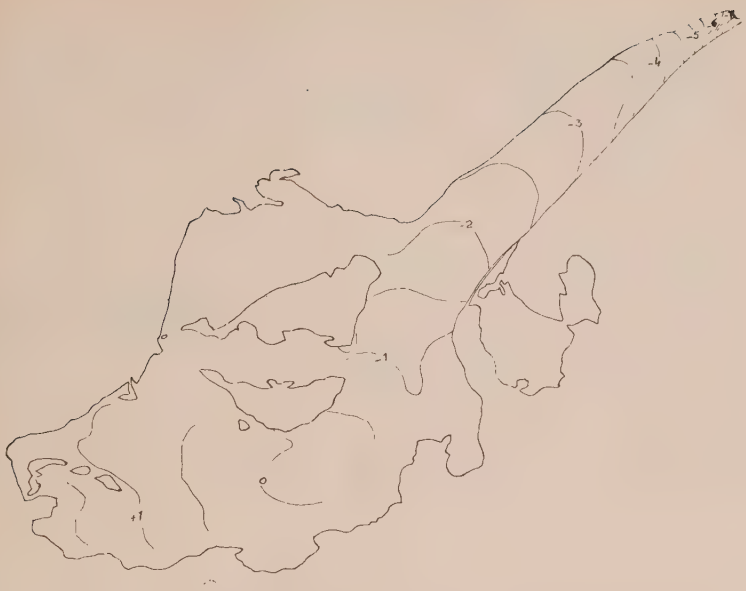
Bereits 1847 wurden durch die Brüder Schlagintweit an der Zunge des Hintereisferners Geschwindigkeitsmessungen durchgeführt (lit. 15).

Im Sommer 1893 begannen A Blümcke und H. Hess eine Untersuchungsreihe, die bis zur Gegenwart praktisch ohne Unterbrechung fortgeführt wurde. Einige ihrer Arbeiten stellen Pioniertaten dar und wären selbst mit den heutigen Hilfsmitteln nur unter größtem Einsatz durchführbar. Das Besondere dieser Arbeiten liegt darin, daß sie, obwohl die intensive Gletscherforschung erst am Beginn war, auf jene Fragen eingestellt waren, die heute noch im Brennpunkt des Interesses stehen.

A. Blümcke und H. Hess haben 1894 mit der Karte des Hintereisferners 1:10000, die eine der ersten Gletscherkarten überhaupt ist, ein wertvolles Dokument geschaffen (lit. 8). Von 1984 bis 1910 wurden eine Reihe von Tiefbohrungen versucht, von denen trotz einfacher Mittel und häufigen widrigen Umständen 12 den Undergrund des Gletschers erreichten. Die Fließgeschwindigkeit wurde an der Zunge mit Hilfe von Steinlinien im Firngebiet an Signalen bestimmt.

1900-1904 untersuchte eine Meßreihe von A. Blümcke und S. Finsterwalder an 18 Pegeln die Unterschiede zwischen Sommer-und Wintergeschwindigkeit. Die vertikale Komponente der Gletscherbewegung wurde estmals gemessen (lit. 6,7).

Nach dem Tode von A. Blümcke im Jahr 1914 wurden die Arbeiten von H. Hess allein weitergeführt. In den Jahren 1917, 1918 und 1919 entstanden kleine Schichtpläne vom Zungenende, die mit der Karte von 1920 den kleinen Gletschervorstoß dieser



Karte 1 — Auftrag und Abschmelzung am Ende des Haushaltsjahres 1952/53 in Metern.

Jahre anschaulich wiedergeben. Die Karte 1:10000 die H. Hess 1920 aufnahm, ist eine der ersten stereophotogrammetrischen Karten überhaupt (lit. 12, 13). Weiteres wertvolles Material lieferten die seismischen Messungen von H. Mothes 1926 und 1928.

Vom Jahre 1932 an übernahmen H. Schatz und Prof. Vietoris die jährlichen Nachmessungen. Ihnen ist es zu danken, daß 1934 im Bereich des Hintereisferners einige Totalisatoren aufgestellt wurden, die seither wertvolle Daten liefern. Auch über die Jahre des zweiten Weltkriges wurden die Beobachtungen fortgesetzt.

Im Rahmen der Arbeitsgemeinschaft arbeiteten im Sommer 1954 H. Hoinkes, N. Untersteiner an Strahlungsbilanzmessungen und Frau Dr. Dirmhirn an Albedomessungen. Mitglieder des Institutes für Geophysik der Universität München führten unter der Leitung von O. Förtsch ausgedehnte seismische Messungen durch.

Im Herbst 1954 übernahm H. Hoinkes einen Teil der Pegel der Arbeitsgemeinschaft für vorbereitende Untersuchungen zum geophysikalischen Jahr. Pläne für weitere Forschungen in den nächsten Jahren lassen hoffen, daß der Hintereisferner auch weiterhin ein Schwerpunkt der ostalpinen Gletscherforschung bleiben wird.

Die Kartographischen Unterlagen:

Die für die Arbeit notwendige Kartengrundlage verdanke ich Prof. R. Finsterwalder. Durch Angehörige seines Institutes wurde im Herbst 1953 der gesamte Hintereisferner photogrammetrisch aufgenommen und im Maßstab 1:10000 ausgewertet. Zusätzliche Aufnahmen der Zunge im Herbst 1952 und im Herbst 1954 ermöglichten, die Stände der Hintereiszunge in allen drei Jahren darzustellen. Das

Gebiet wurde von fünf Standlinien aus aufgenommen. Die Berechnung des Netzes erfolgte im Koordinatensystem der Alpenvereinsvermessung.

Die Karten von A. Blümcke und H. Hess von 1894 und 1920, die Alpenvereinskarte von 1940 und der Schichtplan von 1953 bieten die Möglichkeit über die Volums-, Höhen-und Flächenänderungen des Gletschers genaue Angaben zu gewinnen. Diese Untersuchungen hat bereits Prof. R. Finsterwalder durchgeführt (lit. 10).

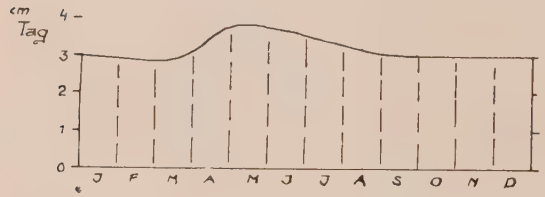


Diagramm 1 — Jahresschwankung der Bewegung. Profil VIII, 3400 m.

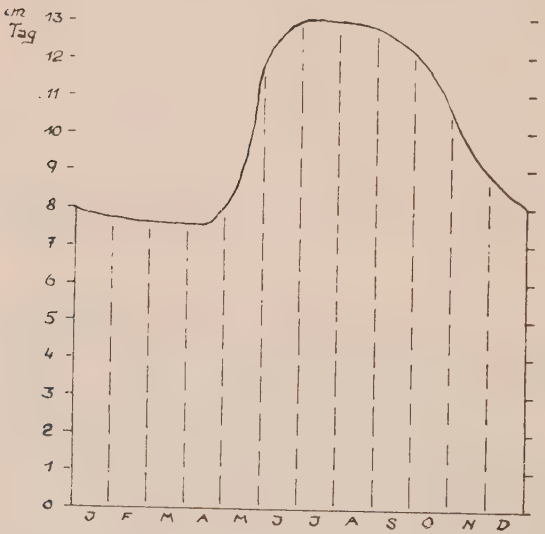


Diagramm 2 — Jahresschwankung der Bewegung. Profil VI, 3000 m.

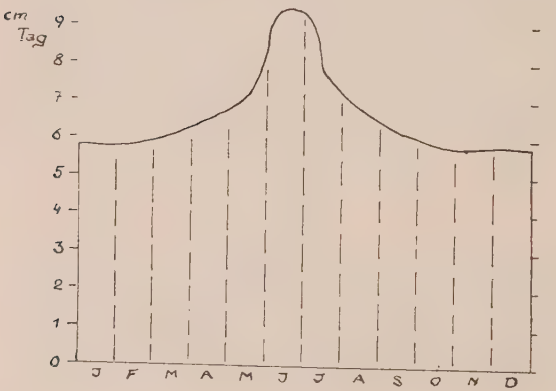


Diagramm 3 — Jahresschwankung der Bewegung. Profil V, 2900 m.

Die horizontale Kompenoente der Gletscherbewegung :

Als eine der bestimmenden Faktoren für die Gletscherdynamik und ihre wechselseitige Beziehung zu den Gletscherschwankungen, war die Gletscherbewegung das wichtigste Untersuchungsobjekt dieser Arbeit.

Aus den vielen Möglichkeiten zur Bestimmung der Fließgeschwindigkeit mußte jene Methode gewählt werden, die einfache Handhabung mit verhältnismäßig großer Genauigkeit verbindet, Diese Bedingungen werden am besten erfüllt, wenn die Bewegung aus Winkeländerungen bestimmt wird.

Um die Fehler aus der oft ungünstigen Lage der Theodolitstandpunkte zum Profil ausschalten zu können, wurden die Winkel auf eine senkrecht zur Fließrichtung stehende Nullrichtung reduziert. Die Entfernung wurde aus den beiden Neigungswinkeln zu den zwei Enden eines 1 m langen, senkrecht stehenden Pegelteiles errechnet.

Ein Vergleich der Durchschnittsgeschwindigkeiten im Längsverlauf des Gletschers zeigt die größte Bewegung nicht im Mittelteil des Gletschers, im Bereich der größten Mächtigkeit des Eises, sondern bereits ober der Firnlinie in einer Höhe von 3000 m. Die Geschwindigkeitsabnahme von der Mitte zum Rand eines Profiles verläuft in einer parabolischen Kurve. Fast im ganzen Gletscherbereich tritt in der Profilmitte eine kleine aber deutliche Geschwindigkeitsabnahme auf. Sie erklärt sich aus den beiden großen Firnbecken und den daraus entstehenden zwei Teilströmen.

Diagramm 4 — Jahresschwankung der Bewegung. Profil III, 2750 m.

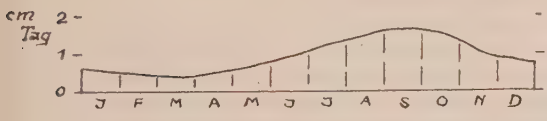


Diagramm 5 — Jahresschwankung der Bewegung. Profil I, 2400 m.

Durch Mitteln der zwei Untersuchungsjahre und durch Ausgleichen grober Sprünge im Bewegungsgang entstanden Schwankungskurven eines Repräsentationsjahres. Der Ausschlag der jahreszeitlichen Schwankung kann 100 % des Jahresmittels betragen. Beim Vergleich dieser Diagramme (Diagramme 1-5) in der Längsache des Gletschers zeichnen sich deutlich starke Verschiebungen ab. Besonders stark wird der Zeitpunkt des Geschwindigkeitsmaximums verlagert. Auch Form und Amplitude der Schwankung ändern sich von Profil zu Profil. Im obe en Firngebiet erhöht sich Ende April, Anfangs Mai die Geschwindigkeit bis zu 25 % über den Jahresdurchschnitt. In den folgenden Monaten nimmt sie langsam wieder ab. In einem Profil bei 3000 m ist die Schwankung breit und mächtig, im Profil 100 m tiefer ist die Geschwindigkeitserhöhung noch stark aber nur mehr einen Monat wirksam. Gletscherabwärts werden die Schwankungen wieder breiter aber auch flacher. An

der Zunge, wo das Bewegungsmaximum in den September fällt, ist der Geschwindigkeitsanstieg flach, der Abfall steiler. In einem Querprofil sind die Jahreskurven der Schwankungen ähnlich. Eine Bewegungsänderung tritt gegenüber dem Stromstrich in den Randbereichen des Profils bis zu einem Monat verspätet auf.

Wenn man für jeden Monat Punkte gleichen Schwankungszustandes in eine Karte einträgt zeigt sich daß die Geschwindigkeitserhöhungen sich wie eine Druckwelle gletscherabwärts fortpflanzen. (Schwankungszutand = zunehmende oder abnehmende Geschwindigkeit, Bewegungsmaximum oder Minimum.) Die im Mai, zur Zeit der höchsten Schneelage, in den oberen Firngebieten entstehenden Zonen von Maxima wandern mit einer Geschwindigkeit von 50-70 m pro Tag gletscherabwärts Die Druckwellen von den kurzen Seitengletschern erreichen den Mittelteil des Gletschers lange bevor die Druckwelle aus dem Hauptfirngebiet in diesen Bereich gelangt. Kurze Zeit laufen zwei Zonen von Maxima parallel gletscherabwärts. Die obere Zone scheint dann ihre Kraft zu verlieren und löst sich im Juli im Bereich der Firnlinie auf. Die untere Welle läuft weiter und erreicht im September die Zunge. Im November bildet sich im Mittelteil des Gletschers eine große Zone von Minima. Sie teilt sich im Jänner, ein Teil wandert ins Firngebiet, der andere zur Zunge.

Ein Vergleich der Ergebnisse dieser Arbeit mit den Werten der Meßreihe von A. Blümcke und S. Finsterwalder (lit. 6) läßt sich nur dann durchführen, wenn man den gewaltigen Gletscherrückgang berücksichtigt. Die Bewegung ist mit dem Rückzug des Gletschers langsamer geworden. Der Gang der jahreszeitlichen Schwankung ist, so weit sich das aus dem damaligen Vergleich zwischen der Sommerbewegung und der Bewegung des übrigen Jahres feststellen läßt, gleich geblieben. Auch die Untersuchungen von W. Pillewizer (lit.14) am Mittelbergferner ergaben ähnliche Verhältnisse.

Die Vergleiche zeigen, daß die Ergebnisse dieser Messung eine verallgemeinernde Verwendung finden können. Der Einbau der gewonnenen Werte in eine Theorie der Gletscherdynamik muß von einem erfahrenen Physiker vorgenommen werden.

Die vertikale Komponenente der Gletscherbewegung :

Die einzige Arbeit, die den vertikalen Anteil der Bewegung untersuchte und auch seine Bedeutung für die Gletscherdynamik berücksichtigte, war die Meßreihe von A. Blümcke und S. Finsterwalder 1900-1904 (lit. 7).

Da die Gletscheroberfläche im Firngebiet und an der oberen Zunge annähernd konstant bleibt, muß die vertikale Bewegungskomponente den größten Teil der Akkumulation und Ablation kompensieren, das heißt im Firnfeld den Betrag des Auftrags abtransportieren und an der Zunge annähernd die Abschemlzung ergänzen.

Die Messung der vertikalen Komponente der Gletscherbewegung brachte nur zum Teil die gewünschten Ergebnisse. Infolge großer Fehler durch die Refraktion war es nicht möglich, die jahreszeitlichen Schwankungen zu untersuchen. Durch Auswahl großer Meßzeiträume, konnte dennoch die für die Haushaltsberechnung notwendige Arbeitsgrundlage gewonnen werden.

Die Berechnung erfolgte aus der Entfernung und dem Tangens des Höhenwinkels. Die Differenz der gewonnenen Höhen ergibt, nach Addition der Korrekturen aus der Instrumentenhöhe und den Veränderungen am Meßpegel, die vertikale Bewegung. Aus der horizontalen und der vertikalen Komponente konnte die Neigung der Bewegung errechnet werden.

In Firngebiet zeigt die Bewegung eine stärkere Neigung nach unten. Gletscherabwärts verläuft die Fließrichtung immer flacher. An den seitlichen Zungenrändern und am Zungenende konnte eine sogar relativ zum Horizont aufwärts gerichtete Bewegung festgestellt werden.

Der tatsächliche Einfluß der vertikalen Bewegung auf den Gletscherhaushtal wird von der Neigung der Bewegung im Bezug auf die Oberfläche des Gletschers gekenntzeichnet. Im Bereich der 2900 m — Höhenlinie verläuft die Bewegung parellel zur Gletscheroberfläche. Nach dem oben gesagten entspricht dies der Firnlinie. Im Firngebiet nimmt die Neigung des Einsinkens nach oben und nach den Seiten hin zu. Gletscherabwärts wird das Aufsteigen der Bewegungslinien aus der Oberfläche immer stärker geneigt.

Auch in diesem Fall ergibt sich eine gute Übereinstimmung mit den Meßwerten von A. Blümcke und S. Finsterwalder. Im oberen Zungenbereich sind die Neigungen sogar absolut gleich geblieben.

Das in gewissem Sinne gesetzmäßige Verhalten der vertikalen Bewegungskomponente ergab eine neue Möglichkeit einer genauen Berechnung des Gletscherhaushaltes. Begründung und Anwendung folgt im Kapitel der Haushaltsberechnung.

Akkumulation und Ablation:

Die Kenntnis des Auftrags und der Abschmelzung, als bestimmende Faktoren des Gletscherhaushaltes, ist eine notwendige Grundlage zur Berechnung der Massenbilanz.

An den 54 Pegelstangen der Geschwindigkeitsmessung wurde auch die Akkumulation oder Ablation gemessen. Durch mehrere Schneedichtemessungen wurde besonders zur Zeit der höchsten Schneelage und am Ende des Haushaltsjahres die Schneedecke ihrer Masse nach bestimmt.

Die Jahresendwerte der Jahre 1952/53 und 1953/54 sind in den Karten 1 und 2 dargestellt. Ein Vergleich der beiden Jahre zeigt, daß bis zu einer Höhe von 2700 m

Karte 2 — Auftrag und Abschmelzung am Ende des Haushaltsjahres 1953/54 in Metern.

die Ablation fast ganz gleich geblieben ist. Die Unterschiede in der Ernährung der zwei Jahre machen sich erst in den oberen Gletschergebieten deutlich bemerkbar. Die Firnlinie liegt im Sommer 1954 100-150 m tiefer und auch der Firnüberschuß ist durchschnittlich 1 m mächtiger als im Vorjahr. Der Unterschied der beiden Jahre ist auch deutlich aus Diagram 6 ersichtlich, das die Kurve aus Ablation und Seehöhe darstellt.

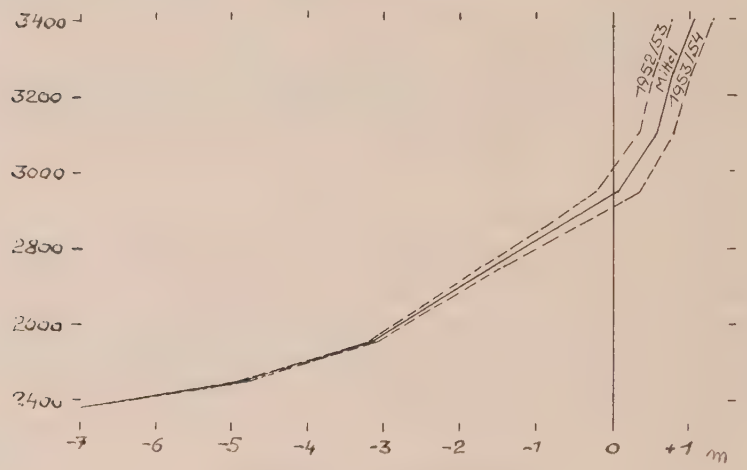


Diagramm 6 — Ablationskurve.

Die Schneedichtemessungen ergaben für die höchste Schneelage im Mai eine durchschnittliche Dichte von 0,35 und für das Ende der Ablationsperiode eine Dichte von 0,5.

Durch Planimetrieren der in Karten eingetragenen Flächen kann man die absoluten Mengen der Akkumulation und Ablation bestimmen. Reduziert man die errechneten Volumen auf die Dichte von Wasser 1,0 und extrapoliert sie auf die gesamte Gletscherfläche, so erhält man für 1952/53 : Auftrag + 1,56 Mill m³, Abschmelzung — 10,20 Mill. m³, Gesamtverlust — 8,84 Mill m³. Für 1953/54 : Auftrag + 3,53 Mill. m³, Abschmelzung —8,49 Mill. m³, Gesamtverlust —4,96 Mill. m³. Das zweite Jahr zeigt also, trotz vollkommen verändertem Ernährungszustand, noch einen großen Verlust in der Massenbilanz. Das Albationsgebiet ist unverhältnismäßig groß. Die Überlänge der Zunge entspricht einem früheren Gletscherstand. Durch eine neue Methode muß also versucht werden, diese Zahlen in ihrer wahren Bedeutung für den Gletscherhaushalt auszuwerten.

Der Haushalt des Hintereisferners in den Jahren 1952/53 und 1953/54 :

Wie schon in der Einleitung gesagt wurde, bietet eine Haushaltsberechnung die Möglichkeit, die Wirkung eines Jahres auf den allgemeinen Ernährungszustand des Gletschers in einer klaren Form zahlenmäßig auszudrücken.

Besonders H. W : son Ahlmann ist es zu danken, daß die Bedeutung einer Haushaltsberechnung für die glaziologische Forschung erkannt wurde. Bei allen seinen Expeditionen (Nordschweden, Spitszbergen und Island ; lit. 1-5) hat er die Bilanzbestimmung zum folgerichtigen Endziel seiner Arbeiten gemacht. Er bestimmt

möglichst genau für den ganzen Gletscher Menge und Dichte des Jahresüberschusses der Akkumulation und die Ablation. Bezieht man die Werte auf die Dichte von Wasser 1,0 und subtrahiert man das abgeschmolzene Volumen vom Volumen der Akkumulation so ergibt sich die Massenbilanz. Diese Methode berücksichtigt jedoch in keiner Weise, daß eine Veränderung im Erhnärungszustand eines Gletschers sich, je nach Größe desselben, erst nach einer kürzeren oder längeren Zeitspanne über den ganzen Gletscher bemerkbar macht. Die Ergebnisse einer solchen Berechnung bringen wohl die absoluten Zahlen der Volumenänderung. Sie können den Zustand des Gletschers nur dann charakterisieren, wenn er seit einer Reihe von Jahren sich zurückzog, stationär blieb oder vorstieß.

Eine Haushaltsberechnung muß daher die verzögerte Auswirkung in den unteren Teilen des Gletschers ausschalten und die Ernährungslage des Firngebietes gesondert betrachten. Das heißt, sie muß das Verhältnis des Jahresüberschusses der Akkumulation zum jährlichen Abfluß aus dem Firngebiet untersuchen.

Die Arbeit von R. Finsterwalder (lit. 9) erfüllt die gestellten Forderungen dadurch, daß er den nach Lagally aus den Oberflächengeschwindigkeiten im Profil der Firnlinie errechneten Eisduchfluß zur Menge des Akkumulationsüberschusses in Beziehung setzt. Trotz der Schwierigkeit nach Lagally halbwegs sichere Werte für den Eisdurchfluß zu berechnen, wurde diese Methode besonders auf Expeditionen mit Erfolg angewandt.

Aus der großen Zahl der oben genannten, verschiedenartigen Messungen am Hintereisferner ergab sich eine neue Möglichkeit zur Berechnung des Gletscherhaushaltes. Diese Methode benützt die vertikale Komponente der Gletscherbewegung. Sie gründet sich auf zwei grundsätzlich Feststellungen:

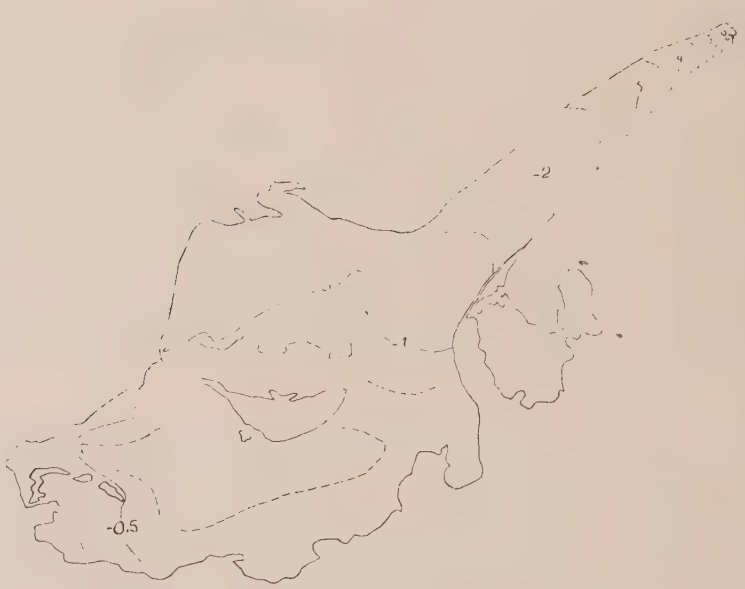
Da an einem stationären Gletscher die Oberfläche von Jahr zu Jahr konstant bleibt, muß zu ihrer Erhaltung der Auftrag oder die Abschmelzung von der vertikalen Bewegung ausgeglichen werden. Im Nährgebiet muß also im Jahr die infolge der vertikalen Bewegungskomponente einsinkende Firnmasse dem Betrag entsprechen den der Jahresüberschuß der Akkumulation bringt. Demgemäß ergänzt im Zehrgebiet das, infolge der vertikalen Bewegungskomponente relativ zur Oberfläche aufsteigende Eis, den Verlust durch die Ablation. Durch Vergleich der Beträge von Auftrag oder Abschmelzung mit den Beträgen der vertikalen Bewegung, kann der Ernährungszustand für jeden einzelnen Punkt des Gletschers, von dem entsprechende Messungen vorliegen, bestimmt werden.

Der folgende Satz behält seine Gültigkeit, auch wenn Auftrag und Abschmelzung, der allgemeine Ernährungszustand des Gletschers und das Größenverhältnis von Firngebiet und Zunge vollkommen außer Acht gelassen werden:

Wenn man die Schmelzung am Untergrund des Gletschers unberücksichtigt läßt, so muß die Volumsänderung im Nährgebiet (bezogen auf die Dichte von Wasser 1,0) durch das jährliche Einsinken gleich gross sein, wie die Volumsänderung (Dichte 1,0) durch das Aufsteigen der Bewegungslinien aus der Oberfläche im Zehrgebiet. Daher entspricht diese Menge auch dem jährlichen Durchfluß (Dichte 1,0) durch das Querprofil der Firnlinie.

Durch Reduzierung der Werte der vertikalen Bewegungskomponente auf die Gletscheroberfläche erhalten wir die durch sie hervorgerufene Höhenänderung (Karte 3). Positives Vorzeichen bedeutet Aufsteigen der Bewegung aus der Oberfläche, negatives bedeutet Einsinken. Addiert man den Jahreswert aus Akkumulation (+) oder Ablation (—) eines Punktes zum Bewegungswert, so erhält man die Bilanz des betreffenden Punktes. Das ist die Höhenänderung der Jahresoberfläche, resultierend aus Bewegung und Auftrag oder Abschmelzung. Ist also der Bilanzwert negativ, so hat sich die Obefläche gesenkt und der Ernährungszustand ist daher negativ. Positive Bilanz bedeutet Hebung der Oberfläche und positiven Ernährungszustand. Diese Werte sind in den Karten 4 und 5 dargestellt.

Karte 3 — Die Höhenänderungen durch die vertikale Bewegungskomponente in Metern.



Karte 4 — Punktweise Jahresbilanz 1952/53 aus Auftrag und Eisabtransport, bzw. Abschmelzung und Eisnachschub (in Metern).

Karte 5 — Punktweise Jahresbilanz 1953/54 aus Auftrag und Eisabtransport, bzw. Abschmelzung und Eisnachschub (in Metern).

Auch hier konnten durch Planimetrieren in den Karten die verschiedenen Volumen berechnet werden. Die Werte wurden auf die Gesamtfläche des Gletschers extrapoliert und auf die Dichte von Wasser 1,0 bezogen.

Besonders die Berechnung der Volumsänderung durch die vertikale Bewegungskomponente brachte gute Ergebnisse. Es ergab sich ein Einsinken im Firnsgebiet von —2,18 Mill. m³ Wasser, das heißt diese Menge wurde auch gletscherabwärts abtransportiert. An der Zunge stiegen durch die vertikale Bewegung Eismengen mit dem Wasserwert von +2,11 Mill. m³ relativ zur Oberfläche auf. Es ergibt sich die unerwartet geringe Differenz von —0,07 Mill. m³. Diese Differenz würde, wenn die Berechnung fehlerfrei wäre, den Eisverlust durch die Druckschmelze und die Erdwärme am Untergrund darstellen. Das wäre ein durchschnittlicher Verlust von —0,75 cm Wasser. Wenn auch in den letzten Jahren der Betrag der Abschmelzung am Untergrund allgemein als sehr klein angenommen wird ist dieser errechnete Wert doch zu gering. Wenn auch die vertikale Bewegung genau gemessen wurde, ergeben sich doch eine Reihe von Fehlerquellen aus der Bestimmung der Oberflächenneigung des Gletschers mit Hilfe der Karte, durch das Interpolieren beim Eintragen und Ungenauigkeiten beim Planimetrieren. Es läßt sich dennoch aus den Werten feststellen, daß trotz der enthaltenen Fehler die oben gemachten Annahmen berechtigt sind. Es ist noch besonders darauf hinzuweisen, daß die errechneten 2,18 Mill. m³ Wasser, die im Firngebiet einsinken, auch dem Durchfluß durch das Profil der Firnlinie entsprechen.

Aus den in den Karten 4 und 5 eingetragenen örtlichen Bilanzen, läßt sich für Nährgebiet und Zehrgebiet eine getrennte Haushaltsberechnung durchführen. Für das Haushaltsjahr 1952/53 ergibt sich auch für das Firngebiet ein Verlust von —3,28 Mill. m³ Firn, das sind —1,64 Mill. m³ Wasser. Die Zunge verlor —8,01 Mill. m³

Eis oder —7,21 Mill. m³ Wasser. Das gibt für den gesamten Gletscher einen Verlust von —8,85 Mill. m³ Wasser, oder als Höhenänderung ausgedrückt, einen durchschnittlichen Höhenverlust von —95 cm. Das Haushaltjahr 1953/54 hat jedoch im Firngebiet eine positive Bilanz mit einem Firnüberschuß von +1,50 Mill m³ oder +0,75 Mill. m³ Wasser. Im Zehrgebiet sind in diesem Jahr —6,92 Mill. m³ Eis, das sind —6,22 Mill. m³ Wasser verloren gegangen. Das ist für den ganzen Gletscher ein Verlust von —5,47 Mill. m³ Wasser oder ein durchschnittlicher Höhenverlust von —59 cm. Auch hier wird, wenn man die Gesamtbilanz des Gletschers betrachtet, der Wert durch die zu lange Zunge und die entsprechend erlahmende Bewegung verfälscht.

Karte 6 — Konstruktion einer dem Ernährungszustand des Gletschers entsprechenden Zunge.

Eine ungefähre Kontrolle der errechneten Werte ergibt sich daraus, daß die Eismasse durch die Bewegung nur verlagert aber nicht verändert wird. Die Gesamtsumme aus Akkumulation und Ablation soll daher mit der Summe der Bilanzberechnung annähernd übereinstimmen. Allerdings ist auch hier die unbekannte Größe der Abschmelzung am Untergrund des Gletschers enthalten. Trotzdem ergibt sich für das Jahr 1952/53 eine erstaunliche Übereinstimmung. Die Summe aus Auftrag und Abschmelzung ergab —8,84 Mill. m³, die Bilanzberechnung —8,85 Mill. m³. Die Fehler durch die ungenaue Bestimmung der Oberflächenneigung und durch das Interpolieren beim Eintragen in die Karte, lassen für das Jahr 1953/54 keine so gute Übereinstimmung zu. Aus Akkumulation und Ablation ergibt sich ein Massenverlust, von —4,96 Mill. m³ und aus der Bilanzberechnung ein Verlust von —5,47 Mill. m³. Ein vollkommen fehlerfreies Arbeiten nach dieser Methode wäre nur mit einem ganz dichten Netz von Meßpegeln und einem riesigen Arbeitsaufwand möglich. Die oben angegebenen gegenseitigen Kontrollen zeigen nur einen maximalen Fehler

von 10 %. Da für eine Haushaltsberechnung auf jeden Fall verschiedene Faktoren nur äußerst schwer zu bestimmen sind, ist dieser Fehler noch tragbar. Diese kontrollierten Ergebnisse beweisen die Richtigkeit der Arbeitshypothese dieser Haushaltsberechnung und zeigen, daß diese Methode auch praktisch anwendbar ist. Die Arbeitsweise dieser Berechnung ermöglicht es also, die Gesamtbilanz so in einzelne Abschnitte aufzuschlüsseln, wie es am Beginn dieses Kapitals gefordert wurde. Diese Methode kann für Nährgebiet und Zehrgebiet, aber bei Bedarf auch für kleinere Gletscherabschnitte, getrennt die Bilanz und damit den Ernährungszustand bestimmen.

Wie bereits erwähnt lassen einige Umstände auf eine Überlänge der Zunge des Hintereisferners schließen. Die oben angeführten Berechnungen geben uns die Möglichkeit eine Zunge zu konstruieren, die dem Ernährungszustand des Firngebietes entspricht. Dafür müssen notwendigerweise drei Dinge bekannt sein, die Durchflußmegen durch das Profil der Firnlinie (aus der Volumsänderung durch die vertikale Bewegung = 2,18 Mill. m³ pro Jahr), der Untergrund des Gletschers (durch die seismischen Messungen von O. Förtsch und H. Vidal; lit. 11) und die Ablation in den verschiedenen Höhenstufen (Diagramm 6). Es wurde auf empirischem Weg eine Zunge mit einer entsprechenden Oberfläche konstruiert, dass die Ablation gleich dem Nachschub von 2,18 Mill. m³ war. Die Firnlinie wurde bei 1950 m angenommen. Es entstand eine theoretische Zunge von nur 2100 m Länge zwischen Firnlinie und Gletscherende. Die heutige Zunge hat aber eine Länge von 5300 m. Noch deutlicher zeigt ein Vergleich der Volumen der wahren und der konstruierten Zunge die Überlänge des Hintereisferners. Während die tatsächliche Zunge 460,8 Mill. m³ Eis beinhaltet, könnte die jetztige Ernährungslage des Gletschers, ensprechend den gegebenen Formen des Untergrundes, nur eine Zunge mit 97,0 Mill. m³ Volumen erhalten. Die Zunge ist also um 363,8 Mill. m³ zu groß. Das Ende der konstruierten Zunge würde direkt unter der Einmündung des Langtauferer-Joch-Ferners liegen, derselbe hätte keine Verbindung mehr mit dem Hauptgletscher. Die konsrtuierte Zunge ist in der Karte 6 dargestellt.

Nach dem Beweis der Richtigkeit und der Verwendbarkeit dieser Methode, soll diese zuletzt angeführte Berechnung zeigen, daß die Anwendung dieser neuen Art der Haushaltsbestimmung berechtigt und notwendig ist.

Zusammenfassung :

Die vorliegende, 1952-54 am Hintereisferner durchgeführte Arbeit hat die Akkumulation, die Ablation und besonders die Gletscherbewegung, in horizontale und vertikale Komponente getrennt, untersucht. Das Endziel der Arbeit war es, diese Ergebnisse zusammenzufassen und daraus den Gletscherhaushalt der beiden Untersuchungsjahre zu bestimmen.

Die Messung der horizontalen Gletscherbewegung brachte besonders hinsichtlich der zeitlichen und räumlichen Verteilung der Jahresschwankungen neue Gesichtspunkte. Die vertikale Komponente zeigte ein gesetzmäßiges Verhalten, im Nährgebiet ein Einsinken relativ zur Oberfläche, im Zehrgebiet ein Aufsteigen. Die Unterschiede der in den zwei Untersuchungsjahren gemessenen Akkumulation und Ablation lassen erkennen, dass die beiden Jahre für die Ernährung des Gletschers engegengesetzte Wirkung hatten und für die jetztigen Verhältnisse Extreme darstellen. Die Haushaltsberechnung mußte mit einer neuen Methode durchgeführt werden. Da die Überlänge der Zunge des Hintereisferners einem früheren Hochstand entspricht, war es notwendig, das Mißverhältnis zwischen Nähr- und Zehrgebiet auszuschalten. Mit Hilfe der vertikalen Bewegungskomponente war es möglich,für Firngebiet und Zunge den Haushalt getrennt zu bestimmen und dadurch die wahren Ernährungsverhältnisse aufzuzeigen. Aus den gegebenen Daten konnte eine Zunge konstruiert werden, die

dem Ernährungszustand des Firngebietes entsprechen würde. Sie hätte nur mehr ein Drittel der Länge der heutigen Zunge.

Die Arbeit konnte durch Anwendung neuer Methoden einen Beitrag zur Lösung verschiedener glaziologischer Fragen leisten, sie hat aber besonders in Bezug auf die Gletscherdynamik einige neue Fragen aufgeworfen.

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L'ACTIVITÉ DU COMITÉ GLACIOLOGIQUE ET LES VARIATIONS DES GLACIERS ITALIENS EN 1956

PROF. MANFREDO VANNI

ABSTRACT

L'auteur rapporte les multiples activités du Comité Glaciologique Italien, s'arrêtant particulièrement sur les séries de données recueillies en trente ans de contrôle des principaux glaciers; enfin, en se basant sur les résultats de la dernière campagne glaciologique de 1956, il montre la possibilité, pour les glaciers alpins, de l'approche d'une nouvelle phase de progression après 40 ans de notable régression.

Le Comité Glaciologique Italien s'occupe activement depuis plusieurs dizaines d'années de l'étude des glaciers.

Créé en 1895 à Turin, grâce à l'initiative du Club Alpin Italien, sous le nom de Commission Glaciologique, il commença à déployer son activité en faisant des recherches variés, de caractère morphologique, descriptif et topographique. Le relief sur grande échelle du Glacier du Miage, sur le Mont Blanc, oeuvre du Prof. F. Porro, date de ces premières années.

Cette Commission intensifia de plus en plus son activité et se transforma en 1914 en Comité Glaciologique Italien, avec comme Président le Professeur Carlo Somigliana, illustre mathématicien et glaciologue de valeur.

Ainsi commença la publication du Bulletin du Comité Glaciologique Italien qui a déjà donné une première série de 25 volumes et une deuxème série de 6 volumes, véritables recueils de travaux importants auxquels ont collaboré l'Institut Géographique Militaire de Florence et le Club Alpin Italien.

Il n'est point de sujets se rapportant à la glaciologie que le Comité Glaciologique n'ait traités et approfondis, des recherches sur le mouvement des glaciers de Somigliana aux études du climat des Alpes de Monterin, en passant par les monographies des grands glaciers du même Monterin, de Vanni, d'Abbadessa, par les études morphologiques de Valbusa, de Capello, de Sacco, par les études plus typiquement géophysiques d'Aliverti, qui se rapportent à la glaciation actuelle, par l'étude du Quaternaire de Sacco, Trevisan, Pasa, etc. qui nous amène à la riche série des relèvements stéreogrammétriques, intensifiés au cours de ces dernières années, et enfin aux recherches hydrologiques et hydrographiques de Volta et de Fantoli.

Outre à ce recueil singulièrement riche, le Comité Glaciologique offre aux hommes d'études les résultats d'une autre de ses importantes activités, c'est-à-dire des données sur les variations des principaux glaciers italiens des Alpes et des Appenins.

Grâce à une campagne continue, faite régulièrement tous les ans depuis 1927, nous avons à notre portée une grande quantité de données extrèmement utiles pour pouvoir déterminer les cycles de variations de la couche glaciaire. C'est sans aucun doute une des plus grands mérites du Comité Glaciologique Italien d'avoir ainsi permis de confronter les variations des glaciers italiens et des glaciers français, suisse et autrichiens.

Ces contrôles réguliers, confrontés avec d'autres documentations sur la glaciation passée, permettent d'arriver à d'intéressantes conclusions sur les cycles glaciaires.

Monterin a pu mettre en évidence qu'il existe probablement un rapport entre les taches solaires qui se forment tous les onze ans et les variations des glaciers, ouvrant ainsi la voie à une série de recherches sur la relation pouvant exister entre le magnétisme solaire, le climat et les dites variations.

Un plus vaste examen permet en outre d'établir qu'il existe sans doute outre les cycles de onze ans, des cycles semi-séculaires, ainsi que l'affirme Köechlin pour les glaciers de la Suisse. En outre, Monterin croit pouvoir démontrer, en reconstruisant des oscillations glaciaires plus amples, qu'il existe également un cycle de 300 ans, qui aurait un rapport avec les variations du niveau marin.

Les résultats que le Comité a atteints sont donc d'un notable intérêt, car ils offrent une série de données qui, jointes à d'autres données plus anciennes, permettent d'établir les cycles des variations glaciaires déjà liés aux cycles climatiques.

Les deux éléments fondamentaux du climat, la température (moyenne estivale) et les précipitations (neigeuses) peuvent être considérés comme les éléments de base permettant d'expliquer et de suivre la marche des cycles glaciaires.

Ainsi que chacun le sait, depuis plusieurs années, et même en tenant compte des oscillations, nos glaciers sont en phase de régression intense et continue. Cet état de choses correspond à une diminution des chutes de neige hivernales et à une légère augmentation de la température au cours des mois allant de Mai à Septembre, selon les données recueillies dans les différentes stations météorologiques alpines.

La hauteur moyenne de la couche de neige hivernale qui oscillait entre les minimums de 1 m 20 et 2 m 20 enregistrés (1) au cours des années 1929/30 et 1936/37, est descendue à des moyennes allant d'un minimum de 20 cm à un maximum de, 1 m 40 durant la période comprise entre les années 1946/47 et 1955/56.

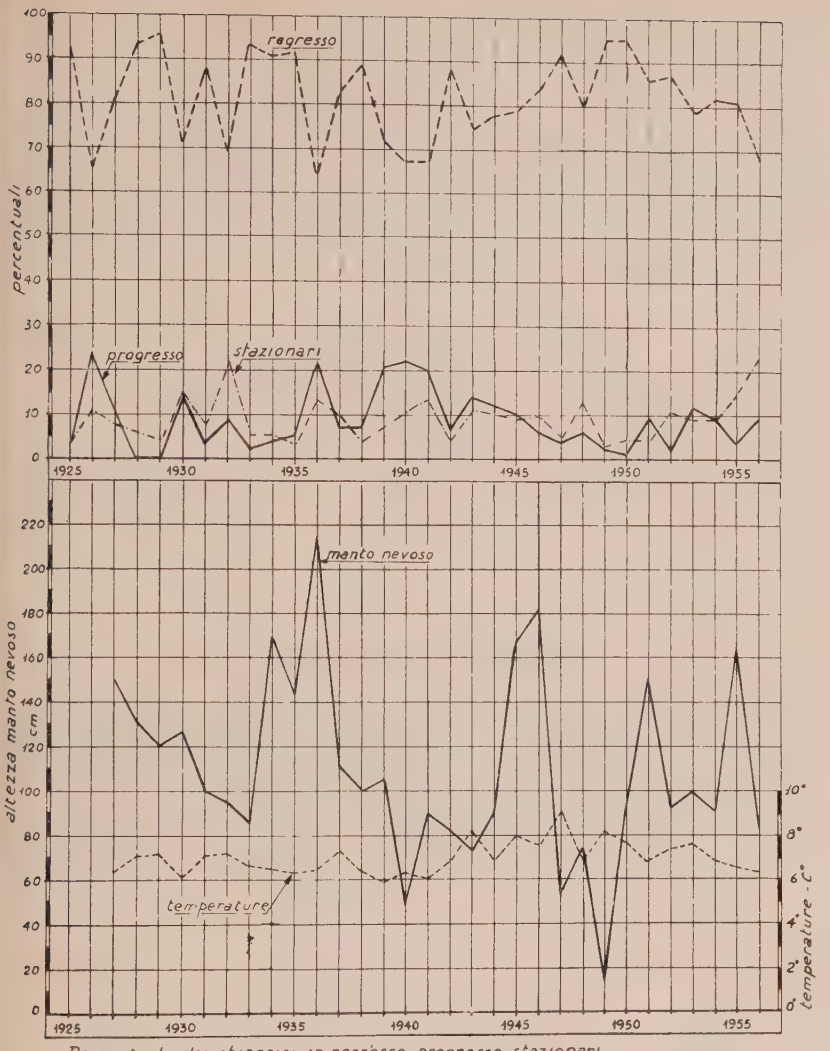
En même temps, les températures moyennes des mois de printemps et d'été qui oscillaient entre un minimum de 4° 5 et 5° 50 durant la période allant de 1929 à 1936, montèrent à 5° et 7° environ entre 1943 et 1956 (2).

Dans l'ensemble donc, depuis de nombreuses années, les 2 facteurs de la glaciation ne sont pas favorables à une conservation et encore moins à une expansion de la couche glaciaire.

L'examen particulier des données recueillies au cours des dernières années et jusqu'à 1956 peut toutefois être intéressant.

En 1956, la campagne glaciologique annuelle s'est déroulée régulièrement; en voici les résultats qui pourront entrer dans le cadre de la marche générale de la glaciation au cours de ces dernières années. 120 glaciers furent contrôlés, dont 9 % se révélèrent en progrès, 23 % stationnaires ou incertains, et 68 % en régression absolument nette et décisive. La phase de régression est donc encore intense en 1956 et est en relation directe avec les conditions météorologiques au cours de l'année. Au cours de l'hiver 1956, c'est-à-dire du mois de Novembre 1955 au mois d'Avril 1956, les chutes de neige n'ont pas été abondantes; la station du Gabiet, par exemple, à 2340 m d'altitude, sur le Mont Rose, a enregistré une hauteur moyenne de la couche de neige de seulement 82 cm, ce qui représente une diminution notable par rapport à la période hivernale précédente au cours de laquelle on était arrivé à 167 cm, mais toutefois supérieure à l'épaisseur enregistrée au cours de l'hiver 1953/54, qui était de 81 cm. La température moyenne des mois de printemps et d'été (Mai à Septembre) fut d'environ 6° 3, inférieure donc à celle de 1936 qui avait été, au cours des mêmes mois, de 6° 5. Le climat de l'année 1955/56 a présenté les mêmes caractères dans les autres stations météorologiques alpines. C'est ainsi qu'à la station du Pian Fedaia, sur la Marmolada, la hauteur de la couche de neige a été de 63 cm par rapport aux 69 cm de l'année 1954/55, et la température moyenne a au contraire présenté une légère augmentation par rapport à 1955, se rapprochant de celles de 1952 et 1953, c'est-à-dire atteignant les 9° 3 et 8° 9, respectivement.

(1) Enregistrées à l'Observatoire du Gabiet (alt. 2340 m) sur le Mont Rose.
(2) Enregistrées à l'Observatoire du Grand St. Bernard (alt. 2467 m) Vallée d'Aoste.



Percentuali dei ghiacciai in regresso-progresso-stazionari
Andamento del manto nevoso e della temperatura estiva al Lago Gabiet (2340 m s.m.)

Si l'on observe les précipitations, on doit toutefois noter que les précipitations de neige hivernales (de Novembre à Avril) ont diminué, mais que les précipitations annuelles ont été plus ou moins les mêmes; en effet, en 1954/55 on a enregistré au Gabiet 777 mm de précipitations, et on en a enregistré 730 mm en 1955/56. Les observations faites en 1956 confirmeraient donc le phénomène qui a été noté depuis diverses années, c'est-à-dire la tendance des précipitations à se déplacer des mois hivernaux aux mois printaniers et estivaux, fait qui influerait sur l'enneigement hivernal et sur la régression glaciaire, étant donné que les neiges d'été ne peuvent alimenter que les bassins montagneux les plus hauts, et n'ont qu'une maigre influence sur les

régions moins élevées, tels que les bassins les plus bas de récolte et les langues glaciaires. Les personnes qui parcourent la haute montagne, au-dessus de 3500 m, peuvent constater combien les glaciers élevés sont neigeux par rapport aux langues découvertes et crevassées, en régression frontale et latérale.

Si l'on examine ensuite les températures des mois de printemps-été (Mai-Septembre) on constate que si, dans l'ensemble, elles sont supérieures à celles des décennies passées, elles sont, par rapport à celles de ces dernières années, en légère diminution, explicable du fait de l'augmentation des précipitations en été et donc du mauvais temps qui favorisent cette diminution, et, en conséquence, minimisent l'ablation des langues glaciaires, ce qui a comme résultat une sensible atténuation de la régression, qui continue, oui, mais au ralenti.

Station du Gabiet (alt. 2340 m)
Température moyennes en C° pour la période Mai-Septembre

1050	7° 6	Noter l'indice graduel
1951	6° 8	de régression de 1950
1952	7° 4	à 1956 (7° 6 à 6° 3 C.)
1953	7° 6	
1954	6° 8	
1955	6° 5	
1956	6° 3	

Les pourcentages des glaciers en régression, stationnaires, incertains, prouvent le phénomène qui est confirmé d'autre part par les observations faites sur les glaciers du versant Suisse, où Mercanton note justement une sensible augmentation des glaciers en phase incertaine et une atténuation de la régression.

Voici un tableau des données concernant nos glaciers, de 1946 à 1956 (1)

	Glaciers en régression (en %)	incertains ou stationnaires	en progrès
1946	84 %	10 %	6 %
1947	92 %	4 %	4 %
1948	80 %	14 %	6 %
1949	95 %	3 %	2 %
1950	95 %	4 %	1 %
1951	86 %	4 %	10 %
1952	87 %	11 %	2 %
1953	79 %	9 %	12 %
1954	82 %	9 %	9 %
1955	80,5 %	15,4 %	4 %
1956	68 %	23 %	9 %

1) — Voir graphique 1

Ces données sont une preuve évidente de l'augmentation notable des glaciers incertains ou stationnaires, ce qui laisse prévoir une phase prochaine de reprise glaciaire, après un ½ siècle d'intense régression, au cours duquel on a pu constater, pour certains glaciers, des reculs de plus de 400 et 500 mètres, et des diminutions d'épaisseur variant entre 40 et 50 mètres.

Les contrôles effectués en 1956 sont en harmonie, dans l'ensemble, avec la marche de ces dix dernières années, c'est-à-dire avec la tendance du phénomène de régression à diminuer, et confirment donc les conclusions des Glaciologues sur les variations des glaciers au cours de l'Histoire, étant donné que les oscillations mineures, ainsi que nous l'avons rappelé, correspondent à des périodes climatiques également mineures, et entrent dans le cadre de cycle glaciaires plus vastes correspondant à leur tour à des cycles climatiques également plus vastes.

Nous sommes donc dans une période de régression apparente, car en observant le phénomène sur plusieurs siècles, nous nous rendons compte que nous faisons partie d'une grande oscillation millénaire positive, que Köechlin appelle *primaire*(3) et que l'on peut représenter par une a grande courbe dans laquelle s'encastrent d'autres courbes mineures ou courbes *secondaires* de 300 ans, et des courbes *semi-séculaires* de 40 à 50 ans, dans lesquelles à leur tour viennent s'encastrer d'autres courbes, les unes *tous les onze ans*, les autres *annuellement*. En conclusion, nous nous trouvons donc dans une période de refroidissement millénaire, mais à un moment de *réchauffement séculaire* : *et c'est ce second phénomène que nous sommes à même de controler et qui joue un role important dans l'histoire contemporaine* (Köechlin).

Les observations faites sur les glaciers italiens confirmeraient que les glaciers ne sont pas destinés à subir d'ultérieures diminutions de leur superficie et de leur épaisseur : dans les siècles à venir, la glaciation devrait s'intesifier graduellement, étant donné qu'une période climatique de refroidissement terrestre, dont la durée devrait être de 1000 à 1500 ans, a commencé au XVIIe siècle.

Nos glaciers, actuellement en régression, sont donc malgré tout bien plus étendus qu'ils ne l'étaient au Moyen-Age, époque à l'aquelle la glaciation vait être notablement plus réduite en raison d'une longue période de climat plus chaud que le notre dont a joui la terre à ce moment-là.

Il apparait donc évident que les contrôles réguliers effectués par le Comité Glaciologique Italien, même s'ils ne portent que sur une période limitée, permettent de tirer des conclusions intéressantes qui pourront être confirmées grâce aux contrôles que, dans l'avenir, ce Comité entend non seulement continuer, mais intensifier.

(3) KOECHLIN R. — Les glaciers et leurs mécanismes — Libraire de l'Université — Lausanne 1944.

MESURES D'ABLATION AU HOFSJOKULL (Islande) 1954

Ch. P. PEGUY
(*France*)

Abstract

Au cours de l'été 1954, une expédition légère française a effectué des observations glaciologiques sur le secteur N. E. du Hofsjökull, coupole glaciaire très surbaissée de quelques 980 kilomètres carrés située à peu près exactement au centre de l'Islande. Il s'agissait des mesures physiques classiques que l'on effectue aux abords de la firn-line.

L'*ablation* est, pour des températures ambiantes identiques, sensiblement plus faible en août qu'en juillet ou juin. Pour l'ensemble de la période de fusion, la perte de substance a été évaluée à une lame d'eau de 1.600 mm à 1000 mètres d'altitude, et de 950 mm à 1.200 mètres.

En s'appuyant d'autre part sur les observations météorologiques les auteurs ont pu fixer, sur ce secteur, le « niveau de *neiges persistantes* » aux environs de 1.150 mètres, le « niveau d'équilibre glaciaire » (notion mécanique intégrant l'hypsométrie du glacier) devant s'établir 60 ou 80 mètres plus haut.

Les diverses langues émises par l'Ice-Cap sur la périphérie présentent des aspects physiques fort variés. Au nord de l'éperon rocheux coté 1.456 mètres, le glacier était, à la mi-août, encore recouvert jusqu'à son front d'une couche de près d'un mètre de neige à gros grains de densité 0,55 à 0,60. Au sud, par contre, de ce même point s'étendait un vaste lobe complètement déneigé, accidenté de bédières et de dirt-cônes. Une hypothèse plausible pour rendre compte d'aspects aussi opposés serait d'imaginer que le névé, à ces altitudes (65° N) s'élabore bien plus lentement que dans les Alpes, si bien que sur les émissaires en pente plus forte, les masses de glace peuvent déjà descendre sensiblement en dessous de la limite climatique des neiges persistantes en conservant un faciès « neige de printemps ».

Au cours de l'été 1954 une expédition légère française a effectué des observations glaciologiques sur le secteur Nord-Est du Hofsjökull, coupole glaciaire très surbaissée de quelques 980 km² située à peu près exactement au centre de l'Islande. Des mesures d'ablation furent notamment effectuées, de la mi-juin à la mi-août, tant sur le glacier lui-même que sur des champs de neige non-permanents situés aux abords du glacier.

1°. Situation topographique.

Les champs de neige non-permanents qui firent l'objet de mesures étaient situés à proximité de la cabane du Laugafell (65°02′ lat. N., 18°18′ W Greenw. alt : 730 m environ) où une station météorologique provisoire fut montée. A cette altitude, la neige couvrait encore, à la mi-juin, plus du quart de la surface du sol, principalement dans les creux où le vent l'avait accumulée. Une telle proportion devait être d'ailleurs, pour l'époque, très anormale par défaut : le bi-mestre avril-mai 1954 avait été en effet, en Islande du Nord, marqué par une température supérieure de plusieurs dégrés aux « normales », ce qui avait dû accélerer la fusion. Moins de deux mois plus tard, d'ailleurs, les derniers vestiges de neige avaient achevé de disparaître, tout au moins en dessous de 850 mètres d'altitude.

En ce qui concerne le glacier lui-même, en notera d'abord que les diveres langues émises par l'Ice-cap sur sa périphérie, même en se cantonnant au secteur Nord-Est que nous avons seul étudié, présentent des apsects physiques très variés.

Au nord de l'éperon rocheux coté 1456 mètres, la pente du glacier est — pour

l'Islande — relativement forte (environ 13 %). Sur cette langue, nous avons trouvé jusqu'à la mi-août, un glacier couvert de neige à gros grains, de densité 0,55 — 0,60, et cela jusqu'à son front, vers 840 mètres. La fusion ne paraissait se manifester — « à vue » — que par des suintements d'eau à la partie supérieure des crevasses : cette eau ne doit pas tarder à se recongeler à nouveau. On n'observait aucune bédière, sauf au voisinage immédiat de la moraine frontale. Par contre, toute une partie du glacier se termine ici sur un lac frontal, non figuré sur la carte au 1/100.000 du Geodetisk Institut, probablement parce que le glacier occupait encore son emplacement lors du levé de celle-ci. La nappe d'eau de ce lac paraît reposer sur de la glace. Dans le lac, le glacier se disloque par rotation en petits icebergs allongés dont la plus grande dimension est parallèle au front.

Dans ces parages, l'expédition installa un thermomètre enregistreur sur une barrette rocheuse du glacier, cote 955 mètres environ.

Au sud de l'épaulement portant la cote 1456 les choses se présentaient de façon toute différente : ici, le glacier descend avec majesté de la zone sommitale elle-même, sur plus de 15 kilomètres, la pente restant de l'ordre de 5 %. On l'abordait par un lobe de glace vive assez crevassé, et il fallait monter aux environs de 1200 mètres pour voir la neige le recouvrir entièrement. Sur ce lobe, on voit l'eau ruisseler partout : les ruisselets se joignent en bédières qui dessinent souvent de véritables méandres. L'approfondissement de ces bédières a été estimé à plus de 10 centimètres par jour en juillet, l'encaissement apparent étant diminué de ce que l'ablation enlève en même temps à la surface même du glacier. La bédière ne peut commencer à se creuser que là où l'érosion gagne de vitesse l'ablation. Si les cours d'eau les plus voisins du front atteignent directement celui-ci, ceux nés dans les parties hautes du glacier se perdent par contre dans les crevasses, disparaissent dans des « moulins », et circulent en profondeur pour résurger peu avant d'atteindre le front. Ces résurgences se font dans des grottes à plancher de glace, il ne paraît pas que nulle part les bédières puissent scier la glace jusqu'à atteindre son substratum.

A sa partie inférieure, le glacier est accidenté de *dirt-cônes* : monticules coniques de sable ou de limon dont la hauteur peut aller de quelques décimètres à plus de 2 mètres. En fait, sables et limons ne forment jamais qu'un matelas de 2 à 3 décimètres d'épaisseur protégeant de la fusion le culot de glace qui forme partout l'ossature du relief : il s'agit donc bien de formes micro-topographiques liées à l'ablation.

On notera enfin que le lobe glaciaire situé au sud de cet éperon 1456 est partout bordé d'un mollisol. Au contraire, au nord de ce point les émissaires du lac frontal, ou de glacier lui-même, traversent un « sandur » bien drainé dès juin.

2°. Les résultats bruts.

Les mesures effectuées comprenaient, répétées à des intervalles de 10-14 jours :

— une lecture directe de l'affleurement de la neige sur une balise.

— un « sondage de battage » permettant d'apprecier le résistanec à la percussion des diverses couches de neige, ou de névé, traversées, jusqu'à 3 mètres de profondeur.

— un « sondage-pesée » permettant de détreminer la valeur en eau de la tranche de neige subsistant au dessus d'un plan-repère.

A. — *Sur un résidu neigeux non-permanent*, un chiffre record d'ablation fut observé les 16-19 juin, vers 720 mètres d'altitude : l'abaissement du niveau lu sur la balise fut de 43 cm en 78 heures, soit environ 13 cm par jour. La densité de la neige étant ici de 0,54 cela correspondait à une valeur en eau de 70 mm. Sans doute cette période vit-elle des températures anormalement élevées (7°3 de température moyenne

pour les 4 jours à notre poste tout voisin) mais une telle ablation ne nous paraît pas moins très exceptionnelle.

Un autre caractère de ces résidus neigeux non-permanents est qu'ils montrent, pour des profondeurs de l'ordre de 50 à 80 cm, une resistance au battage très supérieure à celle offerte, aux mêmes profondeurs, par la neige établie sur substratum glaciaire.

B. — *Sur le glacier* des mesures systématiques furent effectuées sur la langue située *au nord* de la cote 1456, à deux niveau cotant respectivement environ 990 et 1210 mètres. La balise de la cote 990 était située au voisiange du thermomètre enregistreur dont nous avons parlé (environ 35 mètres en altitude au dessus de celui-ci).

Jusqu'au début de juillet l'abaissement de surface du glacier à cette balise inférieure fut en moyenne de l'ordre de 5 cm par jour. Mais l'ablation se réduisit ensuite considérablement en août, sans que la température se soit pour autant déjà nettemene rafraîchie : il faut, pour en rendre compte, invoquer, croyons-nous, l'apparition d'une nuit déjà non négligeable en cette arrière saison.

Les données globales se présentent comme suit :

	Période d'observat.	dQ	dQ/n	A
h = 1210 m	19 VI au 13 VIII	650 mm	12 mm	950 mm
h = 990 m	4 VII au 17 VIII	830 mm	19 mm	1.600 mm

dQ : lame de fusion au cours de la période d'observation.

dQ/n : lame de fusion moyenne par jour.

A : lame de fusion hypothétique pour la période du 1er juin au 30 sept. les valeurs relatives au semaines ayant précédé notre arrivée ou suivi notre départ ayant été calculées d'après les températures du Laugabell, (ou à défaut d'Akureyri).

La limite des *neiges persistantes* paraît s'être établie, pour 1954, dans le secteur étudié, vers 1120-1150 mètres. On peut penser que l'ablation est pratiquement nulle sur la partie la plus élevée du glacier, au dessus de 1550-1600 mètres : au dessus de ce niveau, toute la glace accumulée doit gagner les zones moyennes ou inférieures sous forme solide.

3°. Essai d'interprétation dynamique.

En partant de ces données nécessairement fragmentaires nous avons essayé de supputer un bilan vraisemblable, non de l'ensemble du glacier, mais de ce secteur Nord-Est d'une centaine de kilomètres carrés se terminant par deux langues de part et d'autre de la cote 1456 et qu'il nous a été donné de parcourir.

En combinant la variation de l'ablation en fonction de l'altitude avec la répartition en altitude des surfaces englacées, telle que l'exprime une courbe hyposmétrique (*), nous évaluons l'ordre de grandeur de l'ablation totale aux environs de 95.000.000 m³ d'eau, correspondant sensiblement à une lame d'eau de fusion de 950 mm. Nos observations effectuées à la cote 1210 m paraissent donc significatives des conditions moyennes de ce secteur du glacier; il s'agit d'ailleurs là très sensiblement de l'altitude moyenne même de ce secteur.

Il est beaucoup plus hasardeux d'avancer une estimation de *l'alimentation* du glacier.

L'observation directe du névé, telle qu'on la pratiquerait sur un glacier alpin, ne nous est à peu près ici d'aucun secours. En effet, à la balise inférieure, la « neige à gros grains » que nous avons déjà évoquée, avait une épaisseur de plus de 3 mètres

(*) Détail des calculs et figure dans *Norois*, 1955, art. cité.

à notre arrivée; elle s'avérait remarquablement homogène à l'exception de quelques minces lentilles de glace que la sonde devait traverser; à la mi-août, cette véritable « neige de printemps » *avait* encore une épaisseur dépassant 80 cm, parfois 1 mètre. De telles valeurs paraissent excéder les possibilités de l'alimentation normale à moins de 1.000 mètres d'altitude; nous nous rallierons donc volontiers à l'idée que la neige n'atteint ici la densité du névé, sous ces latitudes, qu'après une évolution de type arctique s'étendant sur plusieurs années.

Signalons toutefois que nous avons plusieurs fois reconnu, au cours de nos « battages », deux couches de neige très molle, tombées sans doute par grand froid, vers 1,60 et 3,00 mètres de profondeur. On peut être tenté de voir dans cette récurrence une indication sur l'ordre de grandeur du « surplus » annuel de l'alimentation sur l'ablation vers 1.000 mètres d'altitude. Divers auteurs ont avancé des estimations de ce surplus pour le Vatnajökull : 2,20 mètres pour l'année budgétaire 1951-1952 au sommet du Bardarbunga (Thorarinsson. Grimsvötn Expédition June-July 1953, pit n° VII, alt 1988 m); 2,40 pour la même année 2,5 km au N de Midfellstindur, NW Oraefajökull, (J. Ives, Nottingham University to Vatnajökull exped. Cité par Thorarinsson, loc, cit, pit n° VIII, alt. 1196 mètres). Des indications de ce genres restent toutefois rares dans les travaux touchant l'Islande. et restent sujettes à caution, car il faut tomber sur des circonstances exceptionnelles, telles que des dépôts de lits de poussières, pour pouvoir identifier dans des névés aussi peu différenciés que les notres des niveaux situés à une année d'intervalle.

On peut enfin recourir au calcul planimétrique pour déterminer l'ordre de grandeur de l'alimentation, ou du moins exclure certains ordres de grandeur inadmissibles. La méthode à suivre consiste à formuler successivement diverses hypothèses relatives au *gradient nivométrique* (accroissement des précipitations neigeuses par 100 mètres d'altitude), pour « cumuler » l'alimentation qu'entrainerait ces diverses hypothèses, compte tenu de l'hypsométrie. Le calcul montre que le gradient qui paraît à tous égards le plus admissible impliquerait sur la zone sommitale des précipitations de l'ordre de 1,50 mètre. Un gradient de cet ordre conduit à un volume de neige reçu annuellement par le glacier qui équilibre sensiblement l'ablation que nous avons calculée à l'instant.

Malgré leur caractère nécessairement fragmentaire, les mesures effectuées permettent ainsi de suggérer déjà deux conclusions :

L'une est qu'il paraît déraisonnable de penser qu'aucun secteur du Hofsjökull puisse recevoir plus de 1,50 ou de 2 mètres de précipitations en année normale, ce qui est relativement peu pour des altitudes dépasant 1700 m. Les mesures de nos confrères que nous avons citées laissent supposer que le versant sud du Vatnajökull offre une alimentation du même ordre dès 1200-1300 mètres d'altitude. Il n'y a pas là de contradiction; le rebord du Hofsjökull que nous avons parcouru appartient à l'Islande Centrale, et celle-ci, sans constituer un domaine réellement aride, est certainement moins puissamment enneigée que la plupart des régions périphériques de l'île.

L'autre est que toutes les méthodes connues d'appréciation de l'altitude des neiges persistantes fixent celles-ci vers 1.100 — 1.150 mètres, *aucune en dessous*. Ce que nous avons dit de l'état du glacier en dessous de cette altitude, au nord de la cote 1456 tout au moins, nous oblige à penser que l'élaboration du névé est assez lente pour que là où la pente du terrain le permet le névé ait le temps de descendre assez bas avant d'avoir atteint la structure et la densité de la glace véritable.

Malgré cette caractéristique remarquable qui l'apparenterait aux glaciers arctiques, notre Hofsjökull ne peu cependant nullement être considéré comme un « micro-inlandsis ». Sur un inlandsis véritable, entre autre, la surface de la zone d'alimentation est proportionnellement beaucoup plus grande — 84 % pour le Groenland —. Pour circonscrire ici une même proportion de la surface englacée

il faudrait descendre jusqu'à l'isohypse de 980 m, ce qui paraît tout de même bien bas, surtout quand on songe à l'état dans lequel nous avons trouvé le lobe du glacier situé au Sud de la cote 1456 m. Il semble donc que ces grands glaciers islandais, tout en présentant certains caractères de détail les apparentant aux glaciers arctiques, conservent les caractères hyspomètriques (et sans doute aussi les caractères thermiques) des glaciers « tempérés ».

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Il pourra être utile de se reporter à la littérature abondante se rapportant au Vatnajökull. Voir la revue *Jökull* (Reykjavik), annuelle depuis 1951.

THE ADVANCE OF THE NISQUALLY GLACIER AT MT. RAINIER, USA, BETWEEN 1952 AND 1956

WALTHER HOFMANN
Munich, Germany

Abstract

The glaciers at the volcanic peaks in the Cascade Range (NW-USA) are advancing since 1950. The advance of the Nisqually Glacier at Mt. Rainier was determined for the interval from 1952-56 by two photogrammetric surveys.

The results are:

1. Linear advance of the front of activ ice: 300 m.
2. Growth in area: 12.57 ha = 2.62% of the 1952-area.
3. Increase in volume: 35 544000 m^3 = 8 886000 m^3/year; Average surface rise: 7.23 m = 1.81 m/year.
4. Increase in speed: The maximum surface speed in a profile crossing the tongue in the elevation 1825 m rose from 90 m/year to 126 m/year which means an increase of 45%.
5. The application of Lagally's formula for the depth of the glacier under the observed speed profile shows that the Nisqually Glacier does not have a laminar flow in its lower parts.

The first permanent glacial advance in this century was observed and reported by american glaciologists from the glaciers at the volcanic peaks in the Cascade Range (State of Washington, northwestern USA) (Lit. (1) and (2)). Retreating in nearly the same rhythm than all other glaciers in the world since 1850, these glaciers begun in 1945 thickening and pushing forward new tongues of activ ice over the old ice bulks, covered with morainal debris.

The best known of these glaciers is the Nisqually Glacier which flows down from the summit ice cape of Mt. Rainier in southern direction. The easy access to this glacier led to continuous observations by the Park Service in the Mt. Rainier National Park during the last 50 years and to several surveys with plane table by A. Johnson from the US Geological Survey since 1945. Whilst the measurements of the Park Service were limited in generally to the observation of the glacier tongue, Johnson measured characteristic cross profiles of the glacier. His measurements gave the first reliable figures of the surface rise and the increase of the Nisqually Glacier. The significance of the event made desirable a survey of the whole glacier by the more effective means of terrestrial photogrammetry. This survey was carried out in the last days of September 1952 by the author with appreciative assistance from the National Park Service and the US Geological Survey.

The photograms was plotted to a map of the whole glacier in the scale 1:10000 with contour lines of 20 meters interval (Lit. (3), (4) and (5)). The survey of the Nisqually was repeated during September 1956 with photograms taken from the same stations and in the same directions like 1952. The evaluation of these photograms in the same way as 4 years before made possible a very accurate determination of the advance in the interval from 1952 to 1956.

The following figures describe the advance of the Nisqually Glacier from different aspects such as :

Linear advance of the activ front
Growth in area
Surface rise and increase in volume
Increase in speed.

1. Linear Advance.

The front of activ ice advanced between 1952 and 1956 the horizontal distance of 300 m. It flowed down over a rocky step in the western part of the glacier canyon and reached the lower part of the canyon which is already filled with remnants of ice, protected by deposits of morainal debris.

2. Growth in Area.

The following figures relate to the area of activ ice of the combined Nisqually and Wilson Glacier. As upper boundary of the Nisqually was taken the 4040 m contour line where it begins obviously as a separate ice stream leaving the summit region of Mr. Rainier. The Wilson Glacier has no connection with the summit ice cape; it begins in a lower basin at an elevation of 3100 m and flows side by side with the Nisqually between 2600 m and 2100 m where both flow together. The lower parts from the 1600 m contour line down to the terminus (elevation 1338 m) are not considered as they are evidently inactiv remnants of a former stand.

Area in 1952:	478.92 ha
Area in 1956:	491.49 ha
Growth:	12.57 ha = 2.62 % of area in 1952
Annual growth:	3.14 ha = 0.66 % of area in 1952

The growth in area of the Nisqually Glacier is not significant for its advance, because its canyon is very narrow, yielding nearly no space at the sides for an increase. 8.67 ha of the overall growth are therefore due to the advance of the *front* of activ ice of the connected Nisqually and Wilson Glacier, 1.25 ha are recently covered, from a partial tongue of the Wilson and only the rest of 2.65 ha are the lateral growth caused by an advance of the lower western margin of the glacier. More significant figures of the glacial increase are therefore given by

3. Surface Rise and Increase of Volume.

The rise of the surface may be seen in diagram 1 which gives a profile along the main axis of advance between 1500 m and 2000 m. The highest rise of 63 m occurred at the new built tongue of activ ice in an elevation of 1600 m. From there the rise decreases to a mean value of 10 m. This value can be observed over the whole Wilson Glacier, whilst the rise decreases considerably at the Nisqually above the confluence with the Wilson. The region higher than 3100 m shows no significant rise.

The increase in volume can be determined by evaluation of the contour maps of 1952 and 1956. A combination of the two maps shows the dislocation of the contour lines. Over the whole Wilson Glacier and the combined Wilson and Nisqually Glacier this dislocation goes in the direction of the glacial flow, indicating the increase. The area G between the same contour line of 1952 and 1956 was considered as the base, the corresponding area g between the next higher contour lines as the cover of a prismatoid, whose volume V was computed with the formula

$$N = \frac{\Delta H}{3}(G + g + \sqrt{Gg}),$$

ΔH signifying the contour interval. As the dislocations are very irregular and different in the lower part of the glacier, the volume of the increase was to compute

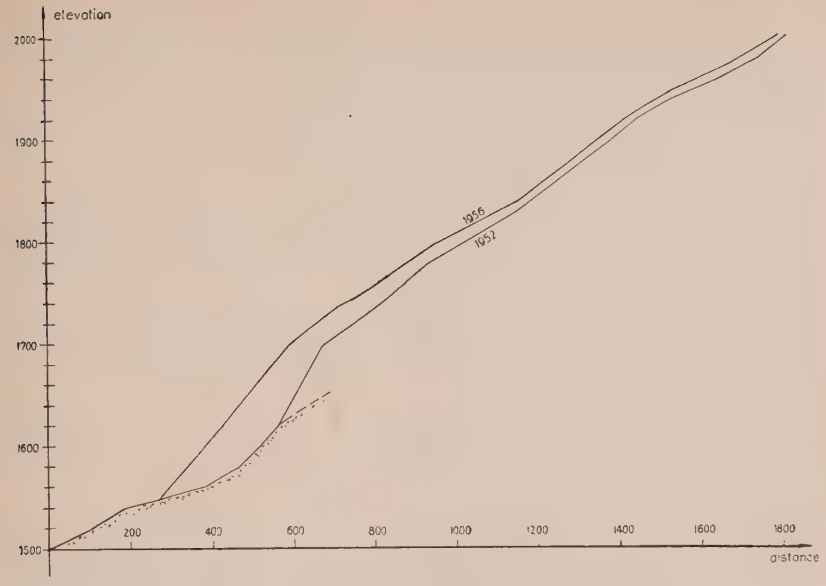

Diagram 1 — Surface rise between 1500 m and 2000 m. Exaggeration of elevation: 2.5 times.

between contour lines of 20 m interval. The upper part, up from 2300 m, allowed the computation from 100 m to 100 m.

The increase in volume, computed in this way, is stated in column 1 of the following table I.

TABLE I

				Wilson Gl			Nisqually Gl		
Zone	V 10^3m^3	A 10^3m^2	R m	V 10^3m^3	A 10^3m^2	R m	V 10^3m^3	A 10^3m^2	R m
1560—1600	834	26.1	31.95						
1600—1700	2340	52.1	44.91						
1700—1800	4208	189.7	22.18						
1800—1900	3368	303.3	11.10						
1900—2000	2988	311.8	9.58						
2000—2100	1582	229.2	5.29						
2100—2200	1772	272.5	6.50	1394	126.8	10.99	364	145.7	2.50
2200—2300	3952	332.6	11 88	3224	244 3	13.20	726	88 3	8 22
2300—2400	3710	292.0	12 71	2780	188 7	14.73	920	103.3	8.91
2400—2500	3040	363.5	8.36	2170	216.7	10.01	870	146.8	5.93
2500—3100	7750	1697.1	4.57	4130	783.3	5.27	3640	913.8	3.98
3100—4040	0	845.0	0	0	0	0	0	845.0	0
Sum	35544	4914.9	7.23	13698	1559.8	8.78	6520	2242.9	2.91
Annual	8886		1.81	3424		2.20	1630		0.73

The total increase in volume appeares as the sum of column 1 with 35 544000 m³.

The average annual increase was 8 886000 m³. These values may be compared with the figures given by V. R. Bender and A. L. Haines in (5), page 280, and corrected to a certain extent by remarks from A. E. Harrison. According to these publications the shrinkage of the Nisqually Glacier between 1910 and 1952 was about 141 millions m³ the most part of which melted between 1920 and 1931 and later between 1936 and 1945, i.e. in about 20 years. The average annual loss was therefore about 7 millions m³. Though these figures are not based on very accurate measurements but depend to a certain extent on estimates, they allow the conclusion that the present advance of the Nisqually Glacier has about the same rhythm than the preceeding retreat since 1920.

As column 1 gives only the absolute increase, these figures are not very fit for cheracterizing the advance. The best measure for such a characterization would be the relation between the increase in volume and the volume itself in a certain zone of the glacier. With the latter, this relation is not available in generally. As a compensation one may consider the relation between the increase in volume V and the surface area A in a certain zone. This relation means the average value of surface rise R in the zone. Column 2 of table I contains the surface areas, the average surface rise is computed in column 3.

From 2100 m upwards, where the Wilson and Nisqually Glaciers can be distinguished, the corresponding values are given for both glaciers separately. These figures show that the advance is considerably higher on the Wilson Glacier whose basin lies in the zone of maximum precipitation. This fact leads to the conclusion that also the increase in the lower part of the Nisqually Glacier comes for the most part from the Wilson Glacier.

4. Increase in Speed.

In 1952, a photogrammetric speed measurement was established in a profile crossing the glacier perpendicularly in the elevation 1825 m. The evaluation of the photograms which covered a period of 4 days is represented in diagram 2. The speed diagram which runs from the left to the right margin of the glacier shows at the left side the normal parabolic increase. But it is quite unusual at the right side. After a

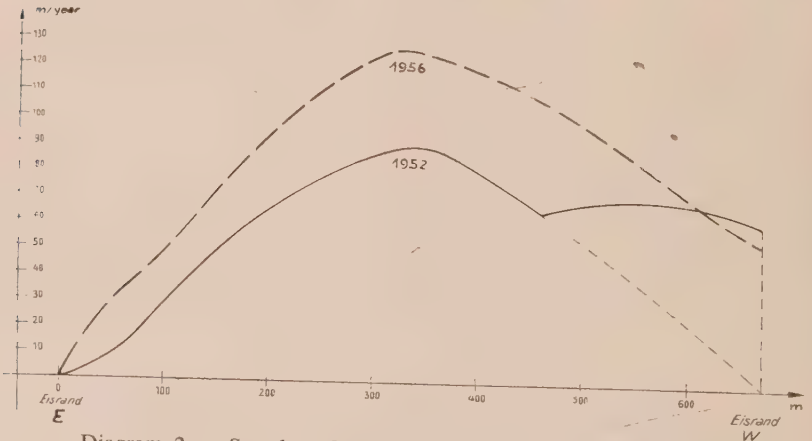

Diagram 2 — Speed profile in elevation 1825 m. 1952 and 1956.

decrease which would bring the speed to zero at the right margin when consequently extended follows a zone of nearly constant speed to the margin itself. This part of the glacier has its origin in the Wilson basin. The course of the speed profile can be interpreted with the fact that the advancing ice bulk of the Wilson Glacier didn't have a close connection with the Nisqually but moved like a separate block over the underlying Nisqually. This fact demonstrates again that the Wilson Glacier was more activ than the Nisqually.

The speed measurement was repeated in 1956 over the same profile. The result is shown likewise in diagram 2 (interrupted line). The speed maximum increased from 90 m/year to 126 m/year which means a growth of 45%. Moreover, the course of the speed profile was normalized in such a way that the crack at the boundary between the Wilson and the Nisqually stream disappeared. Only the relatively high speed at the right margin was preserved.

5. The Glacier Depth and Lagelly's Formula.

In ([6]). Lagally gives a simple formula for the determination of the glacier depth under supposition of a stationary flow and a constant surface speed in a certain profile. The formula can be applied with success also to glaciers with a surface—speed which increases from the margins to the middle as it was shown by R. Finsterwalder (Lit. ([7])). The application is then limited to the centre zone with the maximum speed and the formula gives a good approximation of the maximum depth.
The formula runs:

$$z = \sqrt{\frac{2\,\mu\,u}{\rho . g . \sin\alpha}}$$

wherein means:

z = glacier depth
μ = coefficient of viscosity
ρ = density of glacier ice
g = gravity
α = angle of inclination of the glacier surface
u = surface speed.

The week point in the formula is the coefficient of viscosiy μ which depends on may conditions. It can be determined only by experiments in the glacier ice itself. Lagally gives as a good approximation:

$$\mu = 1.0.10^{14} \text{ g cm}^{-1} \text{ sec}^{-1}.$$

With the values

u = 2.9 10^{-4} cm sec^{-1}
ρ = 0.917 g cm^{-3}
g = 9.80 10^{2} cm sec^{-2}
$\sin\alpha$ = 0.1788

the maximum depth of the Nisqually in the 1952 profile was computed to

$$z = 188.7 \text{ m}.$$

With respect to the morphological conditions of the glacial canyon, this value seams possible.

The repetition of the measurement in 1956 in the same but rised profile give the possibility to compare the real increase in thickness with the difference between the computed depths of 1952 and 1956 and by this to test if the application of Lagally's formula in the case of the Nisqually Glacier is allowed or not.

The new values

$$u = 4.0 \text{ cm sec}^{-1}$$

$$\sin\alpha \quad 0.1521$$

give the maximum depth of

$$z = 241.8 \text{ m.}$$

for the Nisqually profile in 1956.

The difference should correspond to a rise of

$$\Delta z = 53.1 \text{ m,}$$

whilst the real rise, taken from the contour maps is only

$$\Delta h = 13.5 \text{m.}$$

The difference can not be caused only from the uncertainity of the coefficient of viscosity. The computed value of the rise surpasses the observed value about 4 times. As it depends on the square root of the coefficient of viscosity, the latter should be about 16 times smaller than supposed, whilst Lagally's value can be uncertain with a factor of only 2 to 3.

This result shows that the Nisqually Glacier doesn't have a laminar movement in its lower part. The flow is combined with gliding and splintering movements which prohibit the application of Lagally's formula.

The glacial advance extends over all glaciers in the northwestern USA and causes effects similar to those at the Nisqually Glacier. The climatic change which causes the advance may be only a local phenomenon. But the particular structure of the concerned glaciers suggests the supposition that they indicate a worldwide change in climate. With their relatively small bulks and their steep descent, they can more quickly be affected than other glaciers with large firnfields. Further intense observation will therefore be of special importance.

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NOTICE SUR LES VARIATIONS DES GLACIERS DU MONT-BLANC

M. BOUVEROT

Résumé

Le versant français du massif du Mont Blanc se caractérise par des lignes de crêtes élevées (3.600 m à 4.800 m) enserrant des vallées profondes. Les précipitations sont importantes, le climat est froid. Les glaciers ont une surface de 115 km² et leurs langues descendant à proximité des villages.

Les observations ont porté sur des profils en travers et sur les fronts de sept glaciers.

Les graphiques représentant l'évolution des niveaux moyens sur les différents profils sont assez semblables pour chaque glacier : les variations de l'alimentation dans la zone d'accumulation du glacier se répercutent avec un gros décalage et d'une manière progressive sur les différents profils de la langue terminale, tandis que les variations dans l'ablation sont immédiates sur tous les profils.

L'évolution des fronts traduit une décrue presque continue depuis 120 à 150 ans.

Le petit glacier de Tête Rousse a dans son ensemble subi en 55 ans une ablation de 45 % environ.

Les présentes notes ne concernent que la partie française du massif du Mont-Blanc proprement dit, délimitée par les frontières franco-suisse et franco-italienne, et les vallées de l'Arve et du Bon-Nant. Le massif des Aiguilles Rouges et ses petits glaciers en sont exclus.

Cette partie du massif cristallin alpin s'étale du col de Balme, tête de la vallée de l'Arve, au col du Bonhomme, tête de la vallée du Bon-Nant, sur une longueur de 38 km et une largeur de 5 à 11 km.

A l'énorme poussée qui a proté le massif cristallin et sa carapace à une très grande altitude a succèdé un démantèlement intense qui a mis à nu et profondément découpé le noyau granitique si bien que le relief se présente partout sous forme de parois presque verticales dominant des vallées ou des cirques situés à basse altitude. Les différences de niveau sur de faibles distances y sont considérables. Ainsi de l'Aiguille Verte à l'Arve il y a 3030 m de dénivelée sur 5 km, 550 de distance horizontale, des Grandes Jorasses à Chamonix : 2600 m sur 11 km, du Mont-Blanc à l'Arve 3800 m sur 8 km, de l'Aiguille de Trè-la-Tête aux Contamines 2750 m sur 7 km. Aussi, bien que les crêtes se situent presque partout entre 3600 et 4800 m d'altitude, l'altitude moyenne de cette partie du massif n'est, selon Blanchard, que de 2759 m, et un tiers environ serait situé à plus de 3.000 m d'altitude.

Les postes météorologiques installés en 1934 par l'administration des Eaux et Forêts dans les vallées de l'Arve et du Bon-Nant nous permettent d'avoir quelques renseignements sur l'abondance des précipitations au cours des 20 dernières années (1935-1955).

Compte-tenu de l'écran constitué par les montagnes voisines, les fonds de ces vallées enregistrent des totaux confortables : Im,02 à Chamonix (1030 m); 1m,22 à Chamonix-le-Tour (1560 m); 1 m, 35 à Contamines-Montjoie (1170 m). On a tenté d'évaluer le total des précipitations à différentes altitudes sur le massif en échelonnant des nivo-pluviomètres totalisateurs du type Mougin, sur le versant Sud, de la Bérangère à Contamines-Montjoie, et sur le versant Nord, de Tête Rousse aux Houches. Les résultats ont été aberrants, et d'ailleurs l'Electricité de France qui a pris notre succession il y a 10 ans en a constaté, grâce à des visites effectuées en toutes saisons,

un fonctionnement parfois défectueux. Des modifications leur ont été apportées. Il est par suite encore trop tôt pour tirer des conclusions des observations faites. Compte tenu des débits de l'Arve à Chamonix, Blanchard estime cependant que « le versant occidental du Mont-Blanc reçoit par kilomètre carré les précipitations les plus étoffées qui s'abattent sur les Alpes Occidentales ».

Ces précipitations sont réparties sur une moyenne de 120 à 130 jours par an (dans les vallées).

Les coefficients pluviométriques saisonniers sont, pour Chamonix-le-Tour : printemps 0,75, été 1,09 automne 0,97; hiver 1.19, Les précipitations sont donc assez régulièrement réparties suivant les saisons, avec cependant un minimum au printemps, et un maximum en hiver.

Une bonne partie de ces précipitations tombent dans les vallées sous forme de neige. Par hiver il neige 41 jours à Chamonix, 61 jours à Chamonix-le-Tour. Les chutes cumulées donneraient en moyenne par hiver une hauteur totale de neige de 2m,88 à Chamonix, 3m,03 à Contamines et 7m,78 à Chamonix-le-Tour. Blanchard a même relevé pour 37 hivers, à Chamonix-le-Tour, une moyenne de 9m,92 soit près de 10 m, avec deux fois plus de 15 m et jamais moins de 4 m. Nos chiffres sont évidemment influencés par la sécheresse relative qui a marqué quelques-unes des dernières années. Il n'en reste pas moins que dès 1400 m d'altitude les précipitations sont importantes et s'effectuent en grande partie sous forme de neige.

En ce qui concerne les températures, les moyennes annuelles, toujours pour les 20 dernières années, sont de 5°,1 à Chamonix-le-Tour, 6°,3 à Chamonix et 5°,8 à Contamines-Montjoie. En 20 années, il n'y a eu à Chamonix-le-Tour que 2 jours très chauds (+ 30° et au-dessus) et il y en a en moyenne par en 3 à Chamonix et 2 aux Contamines. L'été dans les vallées est donc simplement chaud, sans excès.

L'hiver y est long et rigoureux. Il y a en moyenne à Chamonix 146 jours de gelée par an. Il y en a 173 au Tour et 159 aux Contamines. Ceci se traduit par la persistance du manteau neigeux, même dans les vallées.

Ainsi le massif du Mont-Blanc se caractérise, d'une part, par des lignes de crêtes très élevées enserrant des vallées ou des cirques généralement étroits, et par une altitude moyenne relativement forte, d'autre part, par des précipitations abondantes se produisant surtout en hiver, et par une température généralement froide.

Ces conditions font que les masses énormes de neige reçues par le massif s'y tranforment et s'y conservent sous forme de glace.

Dans tout glacier on distingue une zone d'alimentation et une zone d'ablation. La ligne qui partage les deux zones varie légèrement chaque année en fonction surtout des éléments qui provoquent la fusion. Dans le massif du Mont-Blanc, selon nos observations directes, cette ligne se situe en moyenne aux environs de 2.200 m d'altitude soit par exemple un peu au-dessus des séracs du Géant dans la Vallée Blanche, ou aux séracs de la jonction sous les Grands-Mulets.

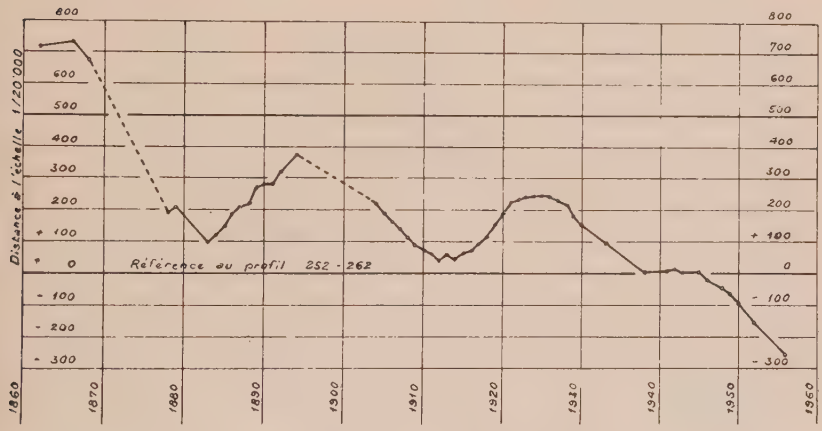
La surface des glaciers est actuellement de 115 km² environ se répartissant ainsi :

Commune de Chamonix	: 88 km2,0
Commune des Houches	: 4 km2,6
Commune de St-Gervais	: 9 km2,2
Commune des Contamines	: 13 km2,2

Ces masses glaciaires poussent leurs langues jusque dans les vallées voisines, près des forêts et non loin des villages. On rencontre ainsi, du Nord au Sud, les glaciers principaux suivants : glaciers du Tour, d'Argentières, de la Mer de Glace, des Bossons, de Bionnassay, de Trè-la-Tête.

Glacier du Tour : Le bassin du Tour, en forme d'amphithéâtre, large et ouvert, est adossé à la frontière franco-suisse-jalonnée par l'Aiguille du Tour (3544 m)

Haute-Savoie. Chamonix. ARGENTIERES
Variation comparée du niveau moyen des profils de 1904 à 1956.



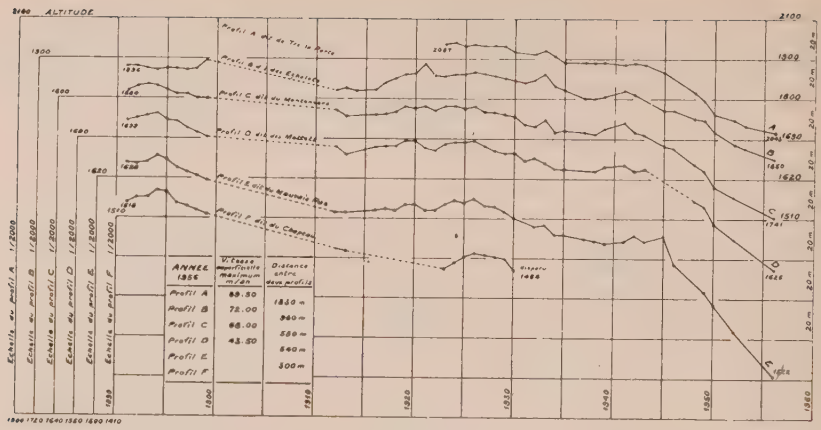
Variation de la langue terminale de 1862 à 1956.

et le Chardonnet (3824 m). Le glacier couvre les pentes supérieures du cirque et se précipite, vers 2200 m d'altitude, sur un escarpement rocheux situé en face du Tour. Sa surface actuelle est de 8 km2,56 dont 6 km2.17 au-dessus de la cote 2800 m.

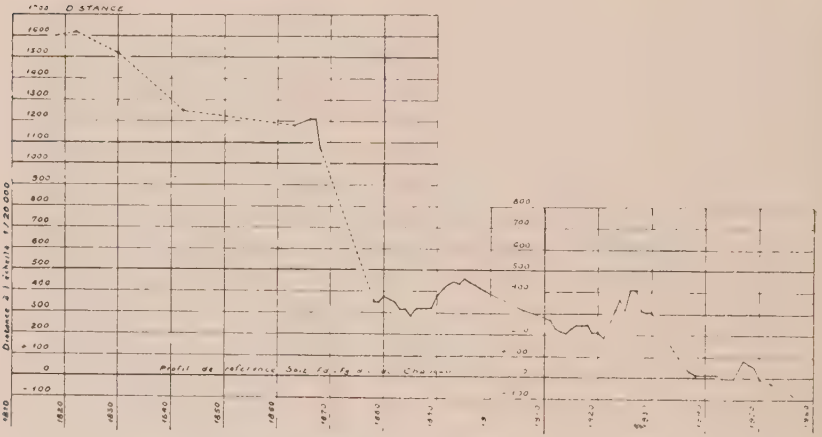
Glacier d'Argentières : De type encaissé, ce glacier suit une vallée de 10 km de longueur dominée par les Aiguilles du Chardonnet (3.824 m), d'Argentières (3.900 m du Mont-Dolent (3.823 m), du Triolet (3.870 m) et de la Verte (4.122 m). Sa surface est de 15 km2,79 dont 9 km2,54 au-dessus de 2.800 m d'altitude.

De la ligne des névés (alt. 2.800 m) à l'extrêmité de la langue, la distance horizontale mesurée suivant l'axe du glacier est de 7 km et la dénivelée de 1360 m.

Glacier de la Mer de Glace : Ce glacier est de type composé. Il est formé par un faisceau de glaciers : glacier de Talèfre dominé par l'Aguille Verte (4.122 m) et le Triolet (3.870 m), glacier de Leschaux dominé par les Grandes Jorasses (4.208 m);



Haute-Savoie. Chamonix. MER DE GLACE
Variation comparée du niveau moyen des profils de 1891 à 1956.

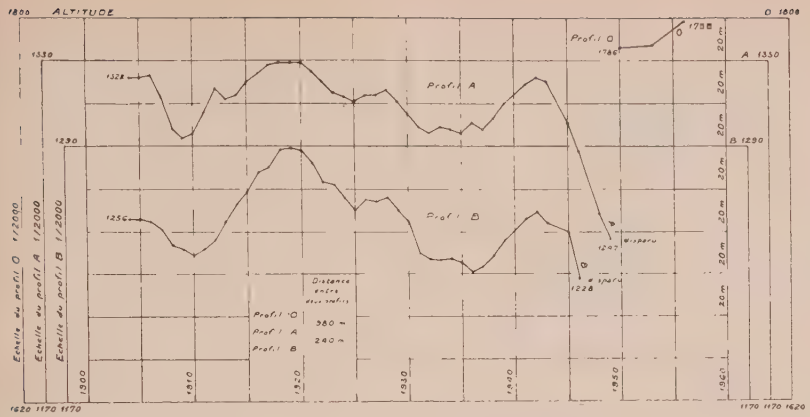


Variation de la langue terminale de 1818 à 1956.

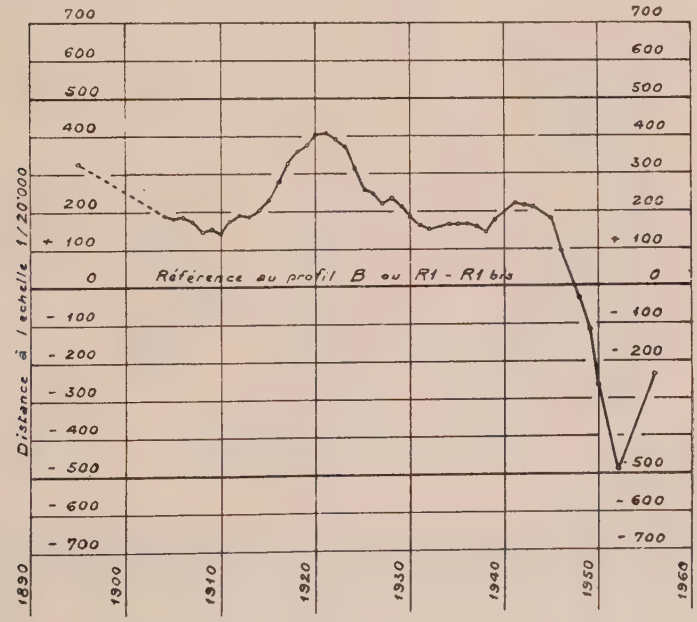
glacier du Géant dominé par l'Aiguille du Géant (4.013 m), la Tour Ronde (3.892 m) et le Mont-Blanc du Tacul (4.228 m), et le Glacier de la Vallée Blanche dominé par l'Aiguille du Midi (3.842 m), la réunion de ces deux glaciers formant le glacier du Tacul. Tous ces glaciers alimentent un tronc commun de 5 km. 200 de longueur, la Mer de Glace, qui s'écoule dans une vallée profonde et encaissée située entre l'Aiguille du Dru (3.754 m) et le Grépon (3.482 m). La surface de cet ensemble glaciaire est de 42 km2,63 dont 27 km2,56 au-dessus de 2.800 m.

De la ligne des névés à l'extrêmité de la langue la distance moyenne calculée sur l'axe des différents glaciers est de 9 km 600, et la dénivelée de 1400 m environ (pente moyenne 7 %).

Glacier des Bossons : Ce glacier descend directement du Mont-Blanc en suivant non pas une vallée mais plutôt un large couloir compris entre le Dôme du Goûter



Haute-Savoie. Chamonix. Les Bossons
Variation comparée du niveau moyen des profils de 1904 à 1956.

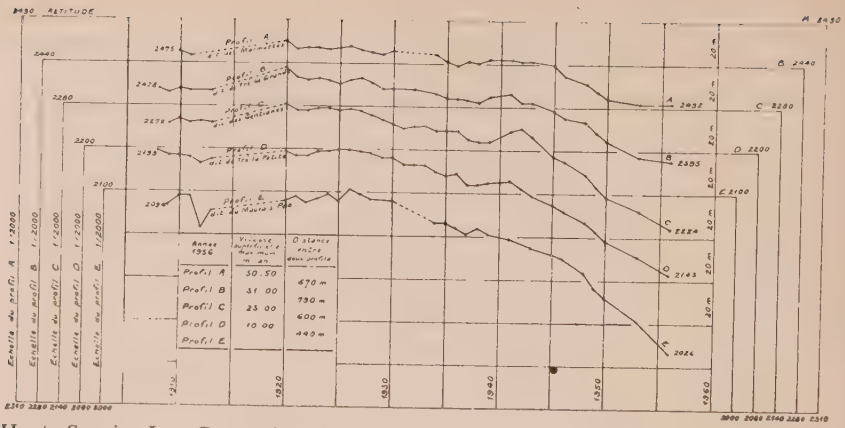


Variation de la langue terminale de 1895 à 1956.

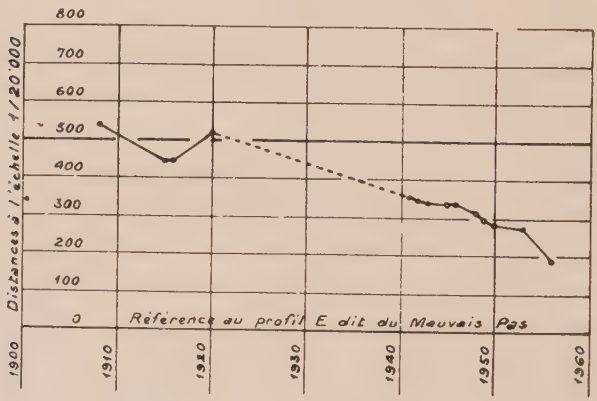
(4.304 m) et les rochers des Grands Mulets d'une part, le Mont-Blanc du Tacul (4.228 m) et l'Aiguille du Midi d'autre part. Sa surface est de 9 km2 au-dessus de 2.800 m et 2 km2, 45 au-dessous, soit en tout 11 km2,45.

De la ligne des névés à l'extrémité de la langue la distance horizontale est de 3 km et la dénivelée de 1560 m (pente moyenne 51 %).

Glacier de Bionnassay : Dominé par les Aiguilles du Goûter (3.863 m) et de Bionnassay (4.051 m) ce glacier occupe une vallée étroite qui débouche sur le Val



Haute-Savoie. Les Contamines Montjoie. TRE LA TETE
Variation comparée du niveau moyen des profils de 1908 à 1956.



Variation de la langue terminale de 1908 à 1956.

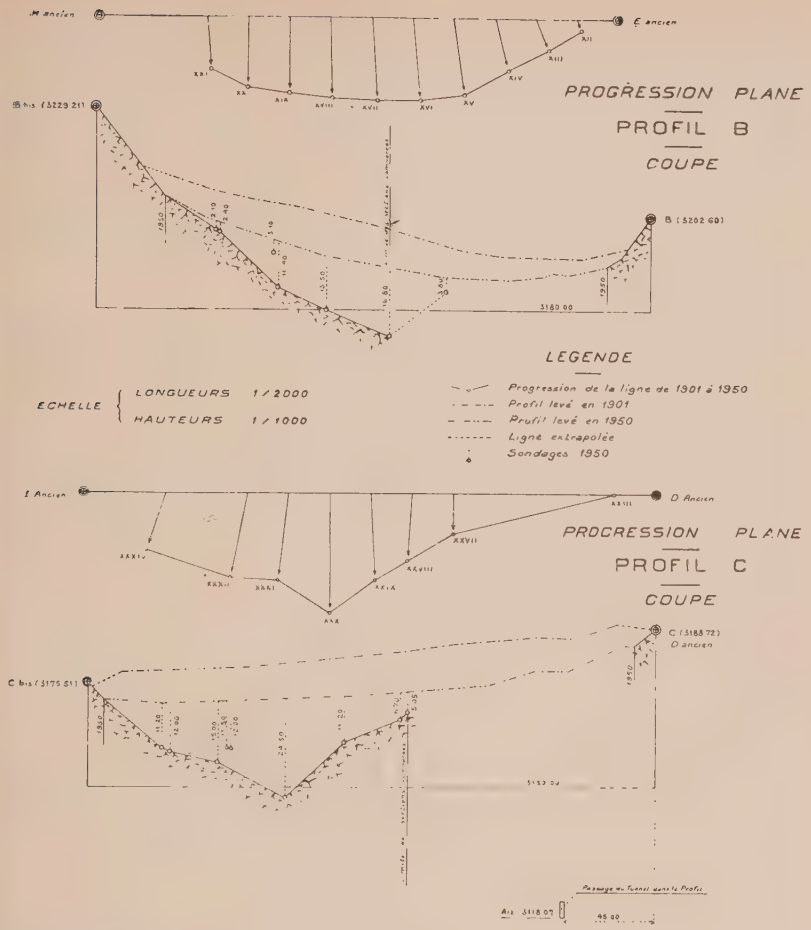
Montjoie et sa langue disparaît sous un amoncellement de matériaux qui forment une moraine frontale. Sa surface est de 4 km,2,37 dont 2 km2,87 au-dessus de 2.800 m.

Glacier de Trè-la-Tête : Ce glacier occupe sur le flanc Sud du Mont-Blanc, près de la frontière franco-italienne une vallée profonde de 7 km,500 de longueur délimitée par l'Aiguille de la Bérangère (3.425 m), le Dôme de Miage (3.670 m), le Col Infranchissable (3.349 m) et les Aiguilles de Trè-la-Tête (3.930 m), des Glaciers (3.816 m) et du Tondu (3.190 m). Sa surface est de 10 km2,62 dont 6 km2,67 au-dessus de 2.800 m.

De la ligne des névés à l'extrêmité de la langue la distance suivant l'axe du glacier est de 4 km,600 et la dénivelée de 870 m (pente poyenne 19 %).

Ces 6 glaciers représentent une surface totale de 93 km2,42 dont 55 km2,26 au-dessus de 2.800 m d'altitude.

En résumé, dans le massif du Mont-Blanc, les bassins d'alimentation sont généralement étendus comparés aux zones d'ablation constituées principalement



Haute-Savoie. Saint Gervais. Tete Rousse
Variations du glacier de 1901 à 1950.

par des langues étroites et allongées descendant très bas dans les vallées. Ces conditions font que les dissipateurs sont doués d'une sensibilité particulière et sont susceptibles de fortes variations.

Observations Glaciologiques

Les études entreprises ont porté sur les langues terminales et ont comporté des relevés de profils en trafers et de front.

Profiles en Travers

On sait que la Mer de Glace, en 1855, remplissait presque complètement ses moraines et était aussi élevée qu'en 1820. Néanmoins les études précises de la langue

terminale, sur profils ,n'ont été faites sur ce glacier que de 1891 à 1899 par M. J. Vallot, et sur la plupart des glaciers par l'administration des Eaux et Forêts depuis 1904 pour les glaciers d'Argentières et des Bossons, 1911 pour la Mer de Glace, 1908 pour Trè-la-Tête. Le relevé d'un profil permet de calculer le niveau moyen de celui-ci au jour du relevé. Nous avons pour chacun des 4 glaciers indiqués ci-dessus, établi des graphiques traduisant pour les langues terminales l'évolution de ces niveaux moyens sur chaque profil.

L'examen de ces graphiques permet de constater que, pour chqaue glacier, les variations des profils sont pratiquement identiques quelle que soit la distance qui sépare ces profils, avec cependant une accentuation plus marquée du mouvement de baisse sur les profils inférieurs.

Les mouvements de crue ou décrue sur les profils sont évidemment la résultante de l'alimentation du glacier dans sa partie haute, et de l'ablation dans la partie basse.

En raison de la distance qui sépare les zones d'alimentation et les profils, et des vitesses différentes dans le différentes parties des sections du glacier, une variation importante dans l'alimentation de la zone supérieure se répercutera sur la langue glaciaire avec un retard énorme de l'ordre de 50 à 100 ans, et s'y trouvera étalée sur plusieurs années. Lorsque la zone d'alimentation comprend plusieurs glaciers d'importance et d'éloignement différents, les variations dans l'alimentation neigeuse des parties hautes retentiront sur le dissipateur d'une manière encore plus progressive. Elles affecteront l'allure générale de l'évolution des profils et le décalage de cette action, profil par profil, sera peu perceptible.

Par contre l'ablation qui est la résultante des condition climatiques sera immédiate et affectera en même temps tous les profils.

Il est donc difficile d'interprèter les renseignements donnés par les variations des niveaux moyens des profils et d'y faire la part revenant à l'alimentation ou à la fusion.

L'étude des variations de niveau, par profils ou simples repères, dans la zone d'alimentation des grands glaciers, encore rarement réalisée, serait donc très utile.

En fait les observations visuelles ou photographiques, quoique non mesurées, permettent de dire que dans la zone d'accumulation la décrue glaciaire se poursuit.

De toute manière sur l'extrêmité des grandes langues glacieaires de la Mer de Glace et de Trè-la-Tête les décrues sont importantes et continues et se sont accélérées depuis 15 ans.

Par contre sur les langues des glaciers d'Argentières et des Bossons, peut-être parce que plus rapprochées de leur source, des alternatives de crue et décrue sont perceptibles avec des maxima en 1904, 1920 et 1942 et des minima en 1910 et 1935-36, le bilan étant en définitive négatif. Toutefois aux Bossons les profils A et B n'étant plus atteints par le glacier, un nouveau profil O installé plus haut en 1950 accuse une crue dans l'intervalle 1953-1956.

Années	Argentières — Profil A			Bossons — Profil A		
	Cotes moyennes	Variations totales	Variations annuelles	Cotes moyennes	Variations totales	Variations annuelles
1904	1866			1322		
		— 9	— 1,5		— 28	— 4,7
1910	1857			1294		
		18	1,8		35	3,5
1920	1875			1329		
					— 33	— 2,2
1935				1296		
					26	3,7
1942		— 58	— 1,6	1322		
					— 75	— 10,7
1949				1247		
1956	1817					
52 ans		— 49		45 ans	— 75	

Années	Mer de Glace Profil de Montenvers			Trè-la-Tête Profil de Trè-la-Petite		
	Cotes moyennes	Variations totales	Variations annuelles[6]	Cotes moyennes	Variations totales	Variations annuelles
1891	1804					
1908		— 29	— 0,6	2199		
					— 13	— 0,4
1941	1775			2186		
		— 34	— 2,3		— 43	— 2,9
1956	1741			2143		
65 ans		— 63	— 1	48 ans	— 56	— 0,8

Vitesses d'écoulement

Les pierres numérotées placées sur les profils, et relevées les années suivantes, permettent de mesurer les vitesses annuelles d'écoulement. Les résultats des obser-

vations de 1956 sur les Glaciers de la Mer de Glace et de Trè-la-Tête sont consignés sur les graphiques joints.

Voici, à titre de comparaison, les vitesses enregistrées par J. Vallot et par nous-même sur la Mer de Glace.

Profil du Montenvers

Année	Cote moyenne	Vitesse annuelle
1892	1806 m	160 m
1899	1800 m	96 m
1956	1741 m	68 m

A Trè-la-Tête les vitesses annuelles en 1956 ont été, dans les différents profils, du haut vers le bas, de 30 m,50; 31 m; 23 m et 10 m.

Mouvement des Fronts

Glacier du Tour — On peut placer en 1643 et 1818 les plus importantes crues du glacier du Tour. En 1818 il atteignait les prairies du hameau du Tour.

Les principaux mouvements enregistrés depuis ont été les suivants (distances horizontales) :

1818 à 1878	— 1.680 m	
1878 à 1892		+ 685 m
1892 à 1908	— 235 m	
1908 à 1920		+ 390
1920 à 1956	— 955 m	
	— 2.870 m	+1075 m

Soit un recul total de 1795 m depuis 1818.

De la cote 1420 environ atteinte en 1818 il est passé en 1956 à la cote 2190 m, soit une remontée de 770 m environ.

Pendant de longues années, la langue s'est trouvée dans une position instable sur le versant du seuil sur lequel débouche actuellement le glacier. Des éboulements spectaculaires de séracs se produisaient fréquemment. Le 14 août 1949, alors que le glacier se trouvait à la cote 2050 environ, des séracs d'un volume approximatif de 450.000 m^3 se sont éboulés, ensevelissant et tuant 6 promeneurs à une distance notable du point de départ.

Glacier d'Argentières — En 1819 il s'avançait jusqu'aux moraines qui longent la rive gauche de l'Arve, et en 1884 il en était à plus de 1.000 m. Une progression s'est ensuite manifestée jusqu'en 1894, suivie d'une régression. Les chiffres suivants indiquent les mouvements enregistrés depuis 1866 :

1866		
	— 631 m	
1883		
		+ 275 m
1896		
	— 329 m	
1912		
		+ 204 m
1925		
	— 503 m	
1956		
	—1463 m	+ 479 m

Soit un recul total de 984 m depuis 1866.
De 1819 à 1956 le front est passé de la cote 1270 (environ) à la cote 1433.

Mer de Glace — A différentes reprises, en 1610 notamment, la Glacier des Bois arriva près du village des Bois, dans le fond de la vallée de l'Arve. En 1825 il était dans un état voisin de son maximum. Son étude fut entreprise en 1846 par Venance Payot, puis en 1890 par le prince Roland Bonaparte, ensuite de 1891 à 1899 par Joseph Payot qui établit les profils des Echelets, du Montenvers, du Mauvais Pas et du Chapeau, enfin depuis 1911 par l'administration des Eaux det Forêts.

Les principaux mouvements du front sont les suivants :

1825		
	— 411 m	
1866		
	— 922 m	
1885		
		+ 174 m
1895		
	— 277 m	
1921		
		+ 169 m
1925		
	— 433 m	
1956		
	—2043 m	+ 343 m

Le front se trouve actuellement à 1700 m du point atteint en 1825. Il est du reste d'observation malaisée parce que situé dans une gorge étroite et inaccessible.

La langue terminale, depuis le profil des Echelets, a eu successivement les surfaces suivantes :

en 1894,	d'après	les	levés	de J. Vallot	149	ha 16	a 58 ca
en 1913,	»	»	»	forestiers	128	72	07
3n 1920,	»	»	»	»	138	84	88
en 1948,	»	»	»	»	92	94	53
en 1956,	»	»	»	»	85	84	56

Glacier des Bossons — Le glacier a eu un maximum vers 1643 puis en 1818. Depuis cette dernière date les mouvements suivants ont été enregistrés :

1818		
	— 682 m	
1874		
		+ 338 m
1894		
	— 190 m	
1910		
		+ 266 m
1920		
	— 262 m	
1938		
		+ 77 m
1941		
	— 708 m	
1953		
		+ 251 m
1956		
	—1842 m	+ 932 m

Soit un recul total de 910 m depuis 1818.

Glacier de Bionnassay — Ce fut probablement vers 1820 qu'eut lieu la dernière crue importante. Depuis, le glacier n'a pratiquement pas cessé de régresser. Le recul de 1644 à 1913 a été de 881 m, et de 40 m depuis 1913. Mais la langue terminale, plus ou moins ensevelie sous la moraine frontale, est difficile à observer.

Glacier de Tré-la-Tête — Compte tenu de la position du front sur la mappe de 1730, le recul du glacier a été pratiquement continu depuis cette date :

1730		
	— 1058 m	
1864		
	— 854 m	
1915		
		+ 74 m
1920		
	— 332 m	
1956		
	— 2244 m	+ 74 m

Soit un recul total de 2170 m depuis 1730.

Le barrage établi sous le glacier en 1944 par l'Electricité de France est maintenant à nu.

Glacier de Tête-Rousse — Une mention spéciale doit être faite sur ce petit glacier suspendu de 12 ha de superficie, situé sur le flanc Est de l'Aiguille du Goûter, entre les glaciers de la Griaz et du Bionnassay, et dont l'altitude varie de 3100 à 3300 m.

En effet le 12 juillet 1892 la rupture d'une poche d'eau dont le volume fut évalué à 200.000 m^3 provoqua dans les vallées de Bionnassay et du Bon-Nant une lave torrentielle qui détruisit une partie des hameaux de Bionnay, du Fayet, et des bains de Saint-Gervais et fit périr 175 personnes.

En 1901 une nouvelle poche d'eau ayant été décelée, une galerie de 208 m dit longueur forée dans le flanc de la montagne, côté Glacier du Bionnassay, perme d'atteindre le fond du glacier de Tête Rousse et d'en drainer les eaux.

Cette galerie continue à être périodiquement entretenue et tout danger de formation de nouvelle poche d'eau semble écarté.

En 1950-1951, à l'aide d'une sonde électrique, nous avons entrepris de dresser sur les profils levés en 1901 des profils sous-glaciaires. Les conditions atmosphériques défavorables, et la présence de nombreux blocs dans le glacier rendirent les travaux difficiles. Les plans ci-joints montrent les résultats obtenus :

Sur la partie du profil C où le rocher en place a été atteint la section du glacier de 1901 à 1951 est passée de 3210 m^2 à 1912 m^2 soit une diminution de 1298 m^2 ou 40 %. Sur la partie correspondante du profil B cette section passe de 1934 m^2 à 959 m^2 soit une diminution de 975 m^2 ou 50 %.

Les pierres peintes placées en 1901 sur les profils et retrouvées en 1950 ont permis de mesurer la vitesse du glacier. Celle-ci a été d'environ 34 m, soit 0 m,70 par an.

GLACIER VARIATION AND TRENDS IN RUN-OFF IN THE CANADIAN CORDILLERA

E. P. COLLIER (*)

The mountainous areas of British Columbia and Alberta are the source regions for the most important water resources of Western Canada. West of the Divide, the water is used mainly for power development. Large parts of the prairies are dependent for irrigation, industrial and domestic supplies and water power upon the run-off from the eastern slopes. Much of Alberta south of Calgary, for example, could not have attained its present development without an assured supply of water for irrigation and domestic use which is sustained by the release of melt water from glaciers.

The variation in glacier movement has been of specific interest to several individuals and agencies in the past and miscellaneous measurements on a few of the well-known Canadian glaciers are available from 1887. In 1945, the Water Resources Branch, Department of Northern Affairs and National Resources, undertook a systematic survey of representative glaciers in the Canadian Cordillera to determine, in a general manner, whether these glaciers were receding or advancing, thickening or ablating at the surface, and varying in forward velocity. Recent heightening of interest in glacier behaviour and a review of the information that has been gathered to date has indicated the need for a reassessment of the adequacy of the present programme.

In this paper, attention is given to the relation between the quantity of water produced by snow and ice melt and the total flow, and to the effect of the glaciers on the régime of the main streams. The paper also presents the results of a preliminary study on trends in run-off in the Canadian Cordillera.

Location of Glaciers

In selecting the glaciers for observation the Water Resources Branch gave preference to those for which previous records had been obtained by other agencies such as the Alpine Club of Canada, and to those which were fairly accessible. Consideration was also given to those glaciers which are the source of rivers with water power possibilities or rivers that constitute a threat to property during floods. In most cases the glaciers are within walking distance of highways or railway lines. Plate I shows the location of the glaciers surveyed by the Water Resources Branch in Alberta and British Columbia.

In the Coast Range of British Columbia, the Helm, Sentinel and Sphinx Glaciers in Garibaldi Park are surveyed. These glaciers lie in the drainage basin of the Cheakamus River. In the Selkirks two glaciers are visited: the Kokanee Glacier which feeds several creeks draining into Kootenay Lake, and the Illecillewaet Glacier which is the source of a tributary of the Columbia River.

In Banff and Jasper National Parks, a survey is made of Peyto Glacier, which is a tongue of the Wapta Icefield and the source of the Mistaya River, a tributary of the North Saskatchewan. Two tongues of the Columbia Icefield are also surveyed:

(*) District Engineer, Water Resources Branch, Department of Northern Affairs and National Resources, Calgary, Alberta.

the Saskatchewan Glacier and the Athabasca Glacier, which are in the headwaters of the North Saskatchewan River and the Athabasca River respectively.

Survey Methods

The surveys are carried out in the month of August, and from 1945 to 1950 inclusive were made annually. Beginning in 1952, the surveys were made every second year. Access is not a problem east of the Continental Divide, but in British Columbia considerably more effort is needed to reach the glaciers, and fresh snow sometimes obscures the ice surface, making it difficult to find markers. The ice front of each glacier is mapped with reference to fixed points on the valley walls to determine its change of position from year to year. Summaries of the results of these surveys for the Peyto, Athabasca and Illecillewaet Glaciers are shown on Plates II, IV and V.

There are certain difficulties in measuring the velocity of surface movement. A transverse line some distance up the snout is carefully related to points on the valley walls; markers or plaques are placed at regular intervals across this transverse profile. Their positions are surveyed on the next visit to the glacier. It has been foundt hat flat metal plates tend to be washed down the glacier by surface streams or be moved about by wind or tourists. Triangular or crossed-strip iron markers have been found to be an improvement. Recently, oxide powder and aniline dyes were tried on the Athabasca Glacier, but with no success. Perhaps the best answer, as suggested by Meier, (see references), is to set wooden dowels a few feet in to the ice, to measure both ablation and forward movement. Tables of the average yearly advance of the

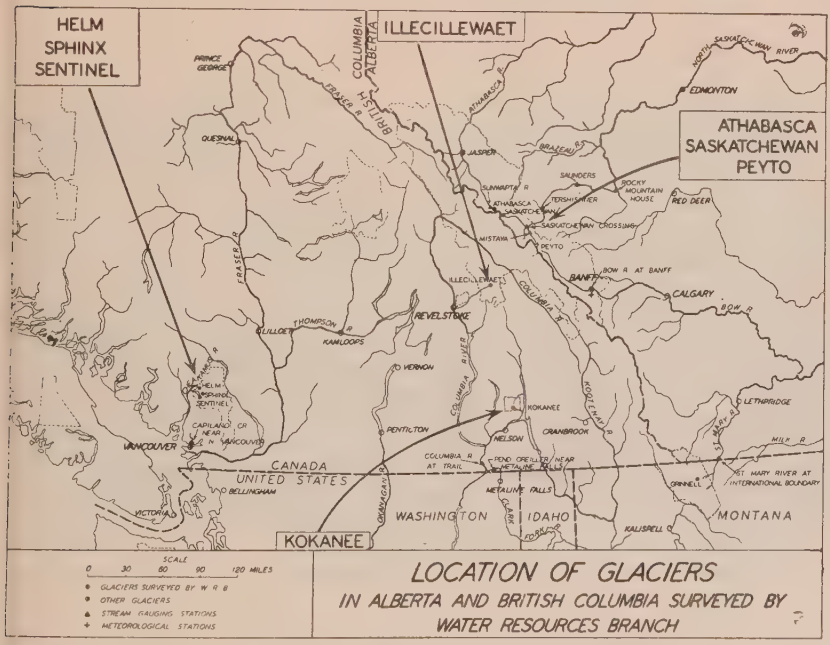


Plate I — Location of Glaciers in Alberta and British Columbia surveyed by the Water Resources Branch.

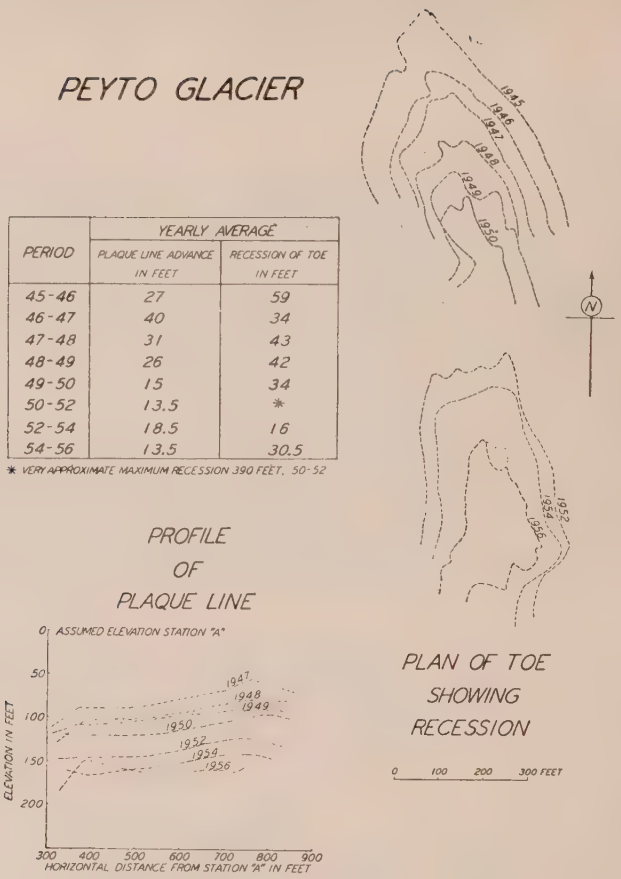


Plate II — Peyto Glacier — Plan of Recession of Toe since 1945, and Transverse Profile at Plaque Line.

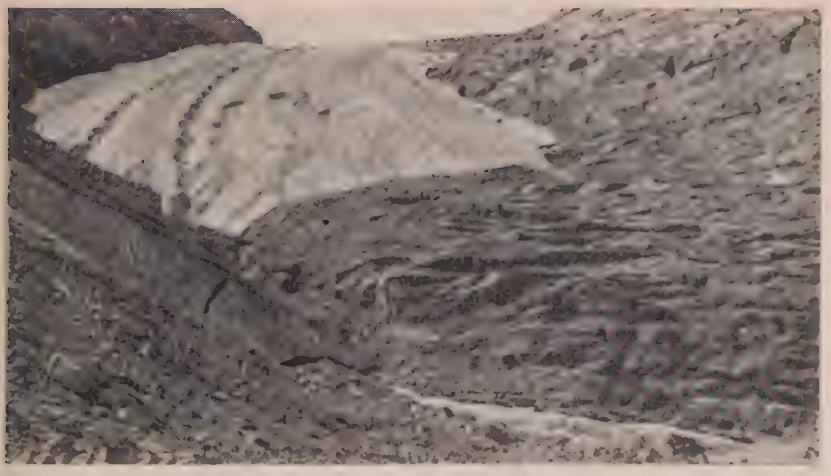
plaque line are shown for the Peyto, Athabasca and Illecillewaet glaciers on Plates II, IV and V. These plates also show transverse profiles of the glaciers at the plaque lines.

Longitudinal profiles are made on some glaciers, usually starting from a point in a transverse profile high up on each glacier and continuing down to the foot, (see Plates IV and V). Photographs from established camera points within the vicinity of the glacier provide a qualitative record of the changes in shape of the tongue. Two examples are shown in Plates III and IV.

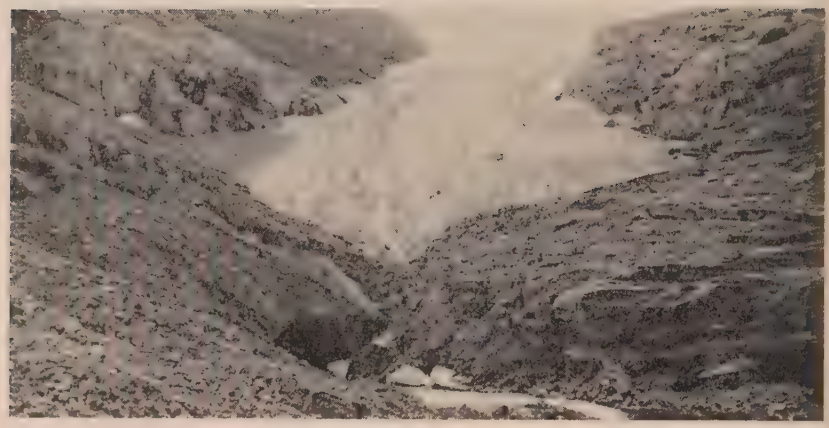
The determination of run-off from ice melt of the glaciers is perhaps the most pertinent aspect of the surveys from the point of view of the Water Resources Branch. Unfortunately, it is often extremely difficult to obtain reliable continuing records of discharge from the glaciers investigated. The stream bed is either extensively braided through moraine, as at the foot of the Saskatchewan, or extremely rough, as at the Peyto. Weather conditions early and late in the season restrict both the length of record that can be obtained and increase the time required to reach the sites. The latter factor is an important consideration in May, when all hydrometric engineers

Plate III — Peyto Glacier — Photographs showing Retreat of Toe.

1950

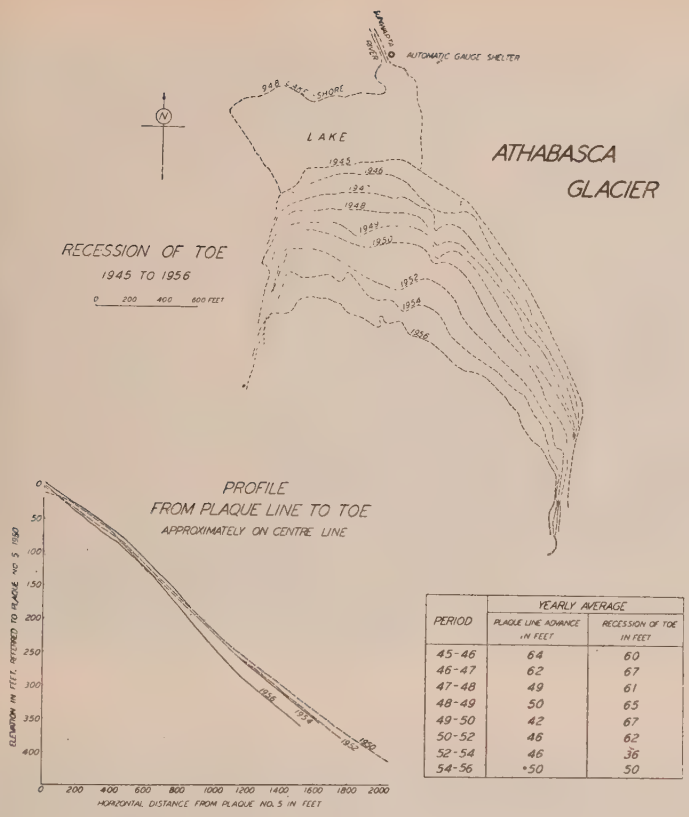


1952



1954

1956



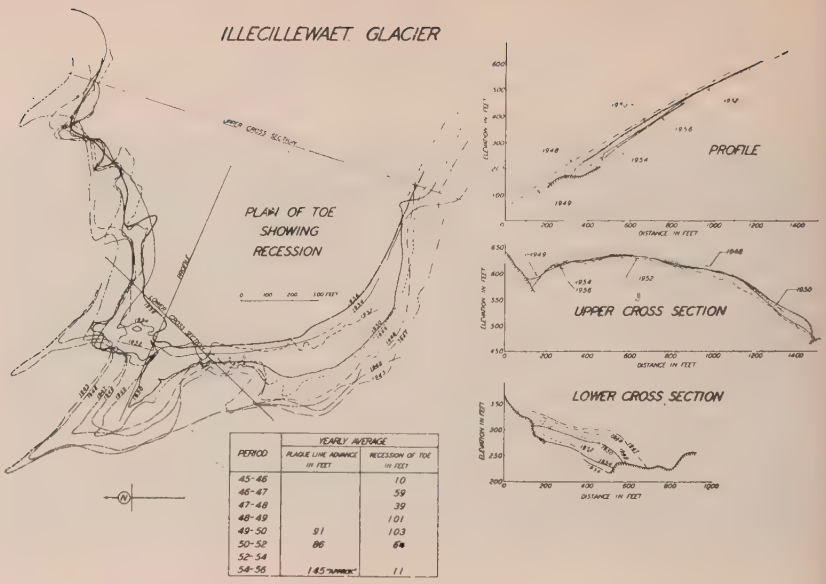
PERIOD	YEARLY AVERAGE	
	PLAQUE LINE ADVANCE IN FEET	RECESSION OF TOE IN FEET
45-46	64	60
46-47	62	67
47-48	49	61
48-49	50	65
49-50	42	67
50-52	46	62
52-54	46	36
54-56	•50	50

Plate IV — Athabasca Glacier — Plan of Recession of Toe since 1945, and Longitudinal Profile from Plaque Line to Toe.

east of the Continental Divide are concentrating on prairie run-off. In only one case, that of the Sunwapta River at Athabasca Glacier, has it been possible to obtain a good location for a stage recorder and measuring section close to the foot of a glacier.

In assessing the effect of glacier variations on water resources, the greatest need is for a more accurate assessment of the volumetric change in the glaciers. Funds are always the limiting factor in extending this type of survey. It is hoped that detail contour maps will be made of the Athabasca tongue in the near future, either by photogrametric or conventional methods, the mapping to be repeated at intervals of perhaps five years. The survey would probably be carried much further up the glacier than the present location of the plaque line.

The results to date confirm the recessive trend which has been noted by other observers in this part of North America for some decades. Some of the most recent surveys, however, show a decrease in the retreat of the ice front and less ablation on the profiles. Observations in Glacier Park, just south of the Canadian border, and elsewhere in the United States suggest that a slight recovery is taking place. Grinnell Glacier in Glacier Park, Montana, for example, appears to have remained in equilibrium for the past five years. It should be noted, however, that whereas most of the



PERIOD	YEARLY AVERAGE	
	PLAQUE LINE ADVANCE IN FEET	RECESSION OF TOE IN FEET
45-46		10
46-47		59
47-48		39
48-49		101
49-50	91	103
50-52	86	64
52-54		
54-56	145 MAX.	11

Plate V — Illecillewaet Glacier — Plan of Recession of Toe since 1945. Transverse and Longitudinal Profiles.

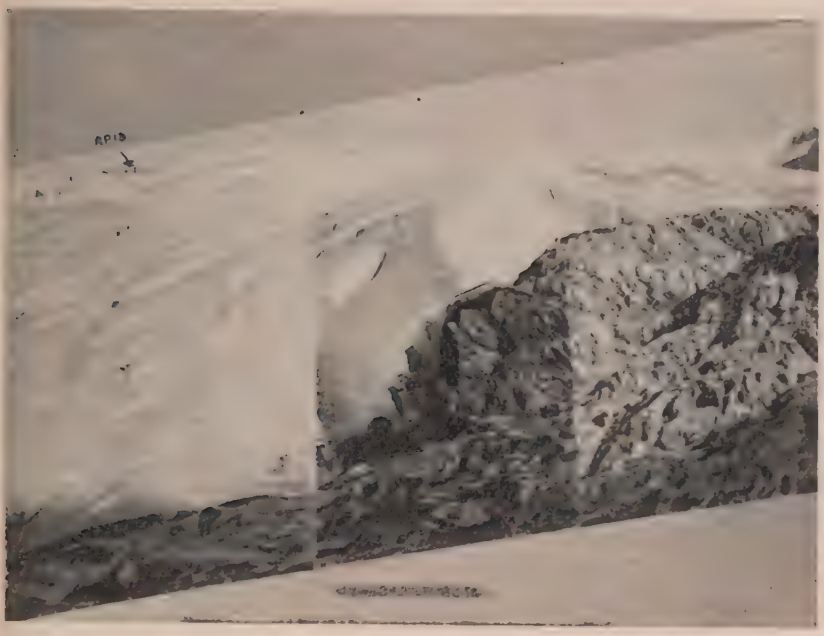
Canadian glaciers are the overflow from large icefields, Grinnell has a very small accumulation field from which to draw its supply. Since the Athabasca and Peyto Glaciers are fed by large icefields, it is reasonable to assume that they would be less sensitive to short-term climatic variations.

Run-Off and its Relation to Glaciers

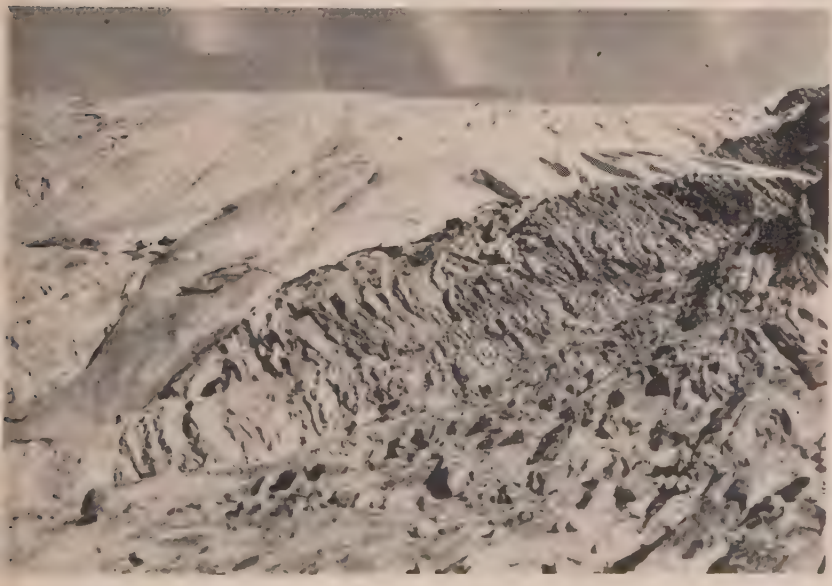
A discussion of the effect of glaciers on run-off generally introduces two main questions: What contribution to stream flow is made by the yearly ice melt of a glacier compared with that made by normal precipitation? What effect has the mountain water on stream régime? At this stage, neither question can be answered with any precision, because the period of record is short and the orientation of the glacier programme has been general. However, it is believed that the results of the surveys by the Water Resources Branch provide the basis for preliminary answers to these questions.

From the positions of the ice front in 1950 and in 1956, from the loss of depth shown by the profiles, and from the forward advance at the plaque line, it is estimated that the Athabasca Glacier yielded, from storage, an average of between 3,000 and 6,000 acre-feet of water per year from ice melt during the period 1950-1956. In making this estimate, it has been assumed that the average velocity of the ice movement is equal to that measured on the surface. Although this assumption may be in appreciable error, it is fortunate that the apparent ablation, which can be more accurately determined from the surveys of the toe and profile, plays a much more significant part in the computation of the total melt. It is interesting to note that A. E. Harrison (see references) estimated a loss in volume averaging about 6,000

Plate VI — Illecillewaet Glacier — Photographs showing Retreat of Toe.



1947



1950

1954

Plate VII — North Saskatchewan River Basin above Edmonton, Alberta, showing Location of Stream Gauging Stations.

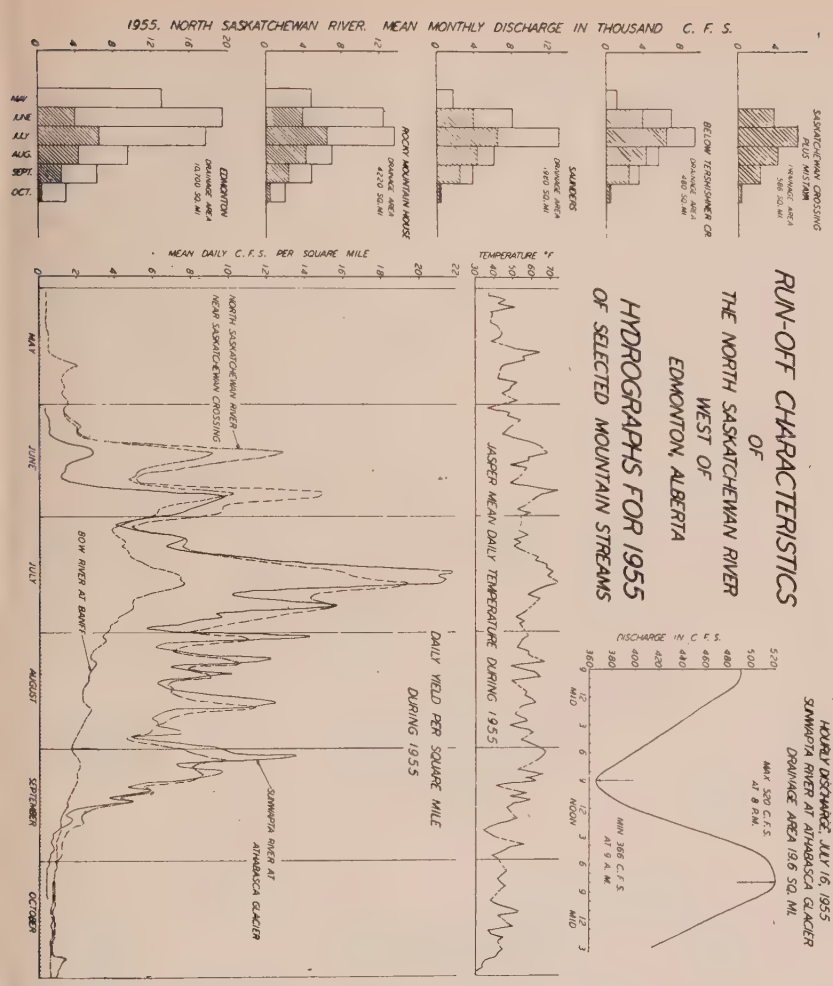

1955. NORTH SASKATCHEWAN RIVER. MEAN MONTHLY DISCHARGE IN THOUSAND C. F. S.
SASKATCHEWAN CROSSING PLUS MISTAWA
BELOW TERRHUMMER CR
SAUNDERS
ROCKY MOUNTAIN HOUSE
EDMONTON
MAY
JUNE
JULY
AUG.
SEPT.
OCT.
MEAN DAILY C.F.S. PER SQUARE MILE
TEMPERATURE °F
NORTH SASKATCHEWAN RIVER NEAR SASKATCHEWAN CROSSING
BOW RIVER AT BANFF
SUNWAPTA RIVER AT ATHABASCA GLACIER
JASPER MEAN DAILY TEMPERATURE DURING 1955
DAILY YIELD PER SQUARE MILE DURING 1955
RUN-OFF CHARACTERISTICS OF THE NORTH SASKATCHEWAN RIVER WEST OF EDMONTON, ALBERTA
HYDROGRAPHS FOR 1955 OF SELECTED MOUNTAIN STREAMS
HOURLY DISCHARGE, JULY 16, 1955 SUNWAPTA RIVER AT ATHABASCA GLACIER DRAINAGE AREA 19.6 SQ. MI.
DISCHARGE IN C.F.S.
MAX. 520 C.F.S. AT 8 P.M.
MIN. 366 C.F.S. AT 9 A.M.
MID
NOON

Plate VIII — Run-off Characteristics of the North Saskatchewan River west of Edmonton, Alberta.
Hydrographs for 1955 of selected mountain streams.

acre-feet per year for twenty years on the Nisqually Glacier on Mount Rainier. Unfortunately, there are no stream flow records from a station at the foot of the Nisqually.

During the same period, 1950-1956, the flow of the Sunwapta River at Athabasca Glacier was approximately 28,500 acre-feet per year. The ice melt, therefore, accounted for about 10% to 20% of the discharge in the stream close to the glacier. These figures appear to modify the popular thesis, which has become familiar on the Canadian prairies, that the wasting away of the glaciers is a serious threat to water supply. The yield from ice melt from Grinnell Glacier is probably also low compared with the total flow in Grinnell Creek.

The effect of elevation on the pattern of the hydrograph of a mountain stream

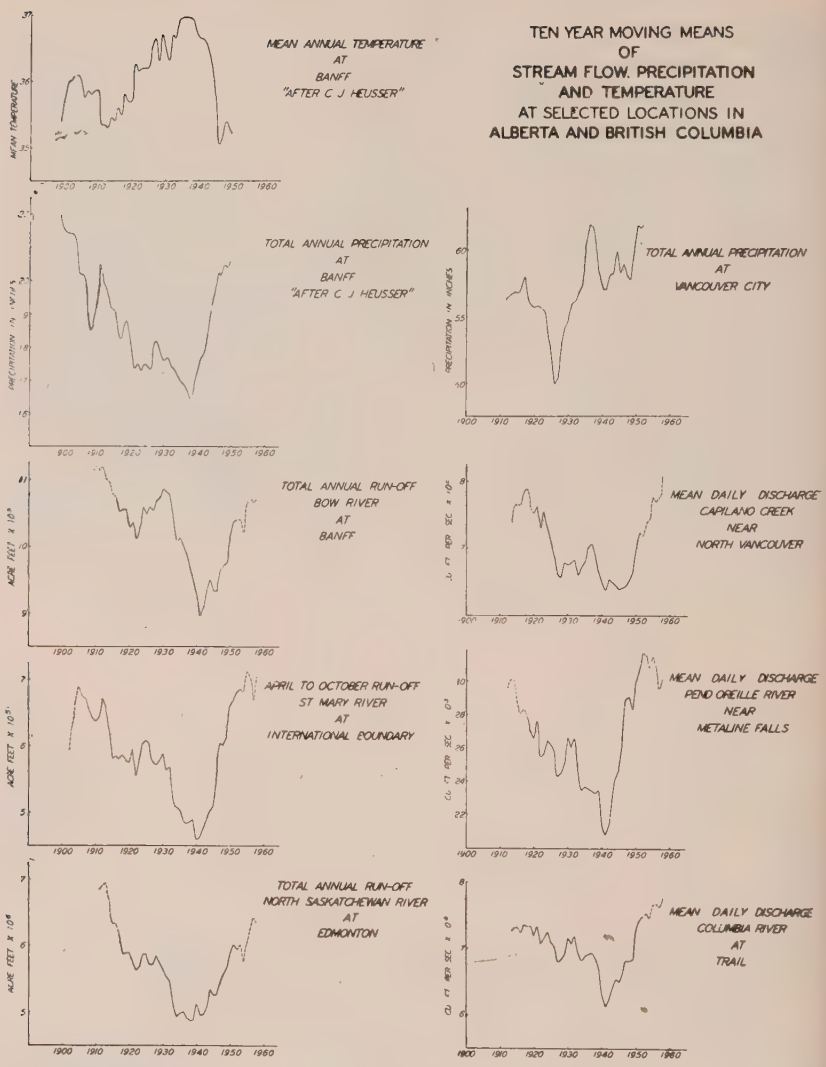


Plate IX — Ten Year Moving Means of Stream Flow, Precipitation and Temperature at selected locations in Alberta and British Columbia.

cannot yet be separated from the effect of ice melt. We do know, however, that the maximum flow from a medium altitude basin (say between 3, 000 ft. and 6,000 ft.) occurs in June in this region, whereas the peak is delayed until July for run-off from higher elevations.

On Plate VIII is shown the 1955 hydrograph of the Bow River at Banff; the peak occurred in late June. This might be termed the valley run-off. Then in July,

a spell of warm weather caused a secondary peak which represented the water from the elevations at which the few glaciers in the Bow basin are located.

On the same plate, the hydrograph for a smaller and higher basin, about 24% of which is covered by ice, is shown to the same scale. The North Saskatchewan River near Saskatchewan Crossing gathers the water at this point from two main mountain tributaries — the Alexandria and the Howse, and itself is fed by the Saskatchewan Glacier and other outflows from the Columbia Icefield. Together these streams draw their supply from a much greater area above an altitude of 6,000 ft. than the Bow system, and collect the drainage from a significant part of the east slope of the icefields along the British Columbia-Alberta boundary. The discharge record is strongly influenced by melt from high elevations and is markedly parallel with the temperature record. Whereas the June peak occurs at about the same time as that on the Bow, the maximum yield occurs in July. Thereafter, until the cool weather in September, the recession from this peak is interrupted during periods of high temperature.

Discharge from the gauging station on the Sunwapta River at the toe of the Athabasca Glacier, is produced by contributions from three sources: relatively low level snow melt on the tongue and surrounding slopes, 6,000 ft. to 8,000 ft.: snow melt from part of Columbia Icefield, 8,000 ft. to 10,000 ft.; and melt water from the ice surface after the snow has been removed. The first June peak is small. In late June the yield does not exceed that of the Bow River or of the Saskatchewan River at the Crossing, but in July the peak yield is even higher than the latter. Other characteristics of the hydrograph are similar, except that the peaks of the Sunwapta River are more pronounced. One interesting point is the steep recession of discharge as soon as the mean daily temperature drops below 50 degrees F. at Jasper. This is to be expected late in the year, since there can be few reserves of snow left below the nevé line, and the air temperature in part reflects the reduction of hours of sunshine.

If data were available to enable a comparison of the yield from a basin without glaciers with that from one with glaciers, e.g., the Sunwapta, there might be appreciable differences. The effect of high altitude plateaux must be to retain a large portion of the snow until late in the season and there may be local climatic effects from the ice itself. There is certainly a delay in the effectiveness of solar radiation in June, due to the unsettled weather in this part of the year along the eastern slopes. These factors, and the presence of the reserve of ice beneath the snow, must work together to produce a later maximum yield and to sustain the yield after run-off is complete in basins without glaciers. Unfortunately, no run-off figures are available for a basin without glaciers and at the same elevation as the Sunwapta basin. Consequently no quantitative comparison can be made to determine the significance of these factors on the annual hydrograph.

Effects on Régime of North Saskatchewan River

A consideration of the flow of the North Saskatchewan River will illustrate qualitatively the effect of the mountain run-off upon the hydrograph at various stations downstream. In 1955, there were five hydrometric stations on the main stem between Edmonton and Saskatchewan Crossing, as shown in the sketch map of the basin. Plate VII. The flow of the North Saskatchewan River at the Crossing is augmented by that of the Mistaya River. The North Saskatchewan Basin has a drainage area of 485 square miles above this point, about 24% of which is covered by ice. The Mistaya River drains an area of 94 square miles above the Crossing, of which about 16% is covered by ice. The total flow is from that portion of the North Saskatchawan drainage system most influenced by high altitude and ice cover.

A typical diurnal fluctuation in discharge at the foot of glaciers is illustrated by the chart for July 16, 1955 for the Sunwapta River (Plate VIII). A similar diurnal

fluctuation is reflected in the daily hydrograph during the summer season in all stations on the main stem of the North Saskatchewan River as far downstream as Edmonton, a distance of about 250 miles and where the drainage area has increased to 10,700 square miles. When the record does not show the wave, the stage may have been smoothed by run-off from a rain storm upstream, or the fluctuation is lost in the flow from snow melt at low levels. The fluctuation is most apparent in late summer when the weather tends to be more settled.

Referring to the diagrams on Plate VIII, it will be noted that the June flow at the Crossing is exceeded in both July and August, with a strongly defined July peak. «Taper out» in October is sharp.

The station on the main stem below Tershishner Creek draws its supply from an area dominated by ranges above 6,000 ft. Flow in July is still the maximum, but the discharge is now greater than that of August. The discharge from above the Crossing exceeds the increment between Saskatchewan Crossing and the gauging site.

At Saunders, which is close to the general 4, 000 ft. contour, about half the June and July supply is accounted for by water from the mountainous area above Saskatchewan Crossing. At Rocky Mountain House, the Clearwater River contributes its run-off. This tributary drains a basin of the same character as that of the main stem between Tershishner Creek and the Crossing. Here the June flow approaches the total of July. Only in August does the Crossing water exceed the additional inflow.

It is at Edmonton that prairie run-off is dominant, despite the inflow from the Brazeau River. There is now a heavy May run-off, a June peak, and strong July flow. In August and September, however, the proportion of mountain water is most effective in sustaining the flow. More than 200 miles downstream, at Frenchman Butte, the mean August 1955 flow had been increased by only 300 c.f.s. for an increase in drainage area of 11,300 square miles. At Prince Albert in Saskatchewan the increase above the Edmonton discharge was 2,400 c.f.s., although the increase in drainage area is 35,400 square miles. The September 1955 run-off figures demonstrate this point in another way. The Edmonton flow was increased by less than 1% for 106% increase in drainage area at Frenchman Butte. The corresponding figures for Prince Albert are 5% for a 331% increase in drainage area.

Trends in Run-Off in Western Canada

Long term climatic trends are apparently reflected in the trends in the annual discharge of streams in mountainous areas, where the drainage basins contain permanent reservoirs in the form of ice and nevé. Unfortunately, stream flow measurements in Western Canada date back only to about the same time as meteorological observations and continuous discharge records of longer than four decades are available for only a very few streams. For instance, the record on the North Saskatchewan River at the Saskatchewan Crossing is a very short one, which is unfortunate because glaciers cover an appreciable proportion of the basin above this point. However, as shown in Plate IX, reference to the longer records of the Bow River at Banff, the St. Mary River at the International Boundary and the North Saskatchewan River at Edmonton indicates a marked parallelism with the curves of temperature and precipitation at Banff as published by C. J. Heusser (see references).

Of particular interest in Heusser's summary was the rise of mean annual temperature at Banff of about 1.7 degrees F. from about 1915 to the late 1930's and the fall of some 2 degrees F. from then to 1950. Corresponding roughly to the same periods, there was a decline in the total annual precipitation of more than 2.5 inches from 1915 to the late 30's and a recovery since then which may be still going on.

The stream flow records illustrate the same trends. At Edmonton, for example, the decline in mean annual flow from 1915 to the late 30's was more than one and a half million acre-feet. An equal recovery has taken place since then. At Banff, the Bow seemed particularly sensitive to the temporary rise in precipitation which occurred in the late 20's. One cannot help but speculate on the fact that the St. Mary River—in the headwaters of which is Grinnell Glacier—has recovered completely to the value of the discharge of the first decade of this century. It will be interesting to learn if the next survey on the glacier shows a positive advance. On the west side of the mountains, stream flows likewise declined during the period 1915 to 1940 and have recovered to at least that of the beginning of the record around 1915. Whether or not this latter trend has reached a peak will be known after a few more years of record.

With the evidence from the glacier surveys by the Water Resources Branch, the indications from the climatic elements and the pattern of stream flow, it is fairly certain that a change from recession to advance of the glaciers of the Western Cordillera is imminent. The advance will be the expression of the recovery from the low of the late thirties. The lag will be of the order of at least 20 years.

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RAPPORT SUR LES VARIATIONS DE LONGUEUR DES GLACIERS EUROPÉENS EN 1953/54, 1954/55 ET 1955/56

PAUL. L. MERCANTON
prof. hon. Lausanne 1957.

Préambule

Le présent rapport, établi pour la commission internationale des Neiges et des Glaces (CING), à l'occasion de la réunion triennale de l'UGGI à Toronto en septembre 1957, résume, comme les précédents, les variations de longueur des glaciers d'Europe en 1953/54, 1954/55 et 1955/56.

On trouvera exposées dans le rapport précédent (publication N° 39 de l'Association internationale d'Hydrologie Scientifique, session de Rome 1954), la genèse et l'histoire de cette étude systématique des variations glaciaires depuis ses débuts en 1880-81.

Il faut toutefois compléter cet historique par le nom du dernier président de l'ancienne commission A. Hoel et de ses membres titulaires pour les diverses contrées jusqu'en 1934; Messieurs :

Antarctique : F. D. Debenham, Scott Polar Institut, Cambridge, Angleterre
Allemagne : H. Hess, prof. Nürnberg.
Autriche : R. von Klebelsberg, prof. Insnbrück, Tyrol.
Canada : A-O. Wheeler, Sidney, Vancouver Island.
Groenland : Lauge Koch, Grönlands Geologisk Undersoegelser, Copenhague.
Islande : J. Eythorsson, Reykjavik.
Etats Unis d'Amérique : Fr. E. Matthes, Washington D.C.
France : Ch. Rabot, Paris.
Géorgie du Sud : O. Holtedahl, Geolog. Institut, Oslo, Norvège.
Grande Bretagne : J. M. Wordie, Cambridge.
Indes Himalaya etc. : Geological Survey of India, Calcutta.
Italie : C. Somigliana, prof. Université, Torino.
Nouvelle-Zélande : A. P. Harper, Wellington.
Norvège : W. Werenskioeld, Geogr. Institut, Oslo.
Russie (U.R.S.S.) : J. Schokalsky, Leningrad.
Suède : H. Ahlmann, prof. Université, Stockholm.
Suisse : P. L. Mercanton, prof. hon. Lausanne.
Svalbard : A. Hoel, Oslo.

Voici maintenant la liste de nos informateurs nationaux actuels, auxquels la Commission des Neiges et des Glaces, ainsi que le rapporteur, adressent leurs vifs remerciements, Messieurs :

Autriche : Prof. Dr. von Klebelsberg, Insbrück.
France : Inspecteur général Messines du Sourbier, pour l'administration des Eaux et Forêts Grenoble.
Italie : Prof. Manfredo Vanni, Turin, pour le Comitato Glaciologico italiano.
Islande : Dr. John Eythorsson, Reykjavik.
Norvège : Dr. Liestoel, Norsk Polar Institute, Oslo.
Suède : Dr. Erik Woknerud, Geografiska Institutet, Stockholm.
Suisse : A. Renaud pour la Commission helvétique des Glaciers, Lausanne.

Ces correspondants nationaux gardent comme toujours la responsabilité de leurs données.

Une récapitulation générale termine le Rapport. On y trouvera pour chacunes des grandes régions englaciées d'Europe, et pour les trois périodes 1953/54, 1954/55 et 1955/56, outre les nombreux totaux d'appareils observés, les nombres des glaciers en crue, stationnaires ou en décrue.

Commission des Neiges et des Glaces,
le Rapporteur : P. L. Mercanton.—
Lausanne, printemps 1957.—

Le Président :
R. Haefeli (Zurich)

Avec nos remerciements très vifs à nos correspondants pour leur information bénévole, laborieuse et délicate.

VARIATIONS DE LONGUEURS DES GLACIERS FRANÇAIS, EN MÈTRES.—

Le signe — signifie : décrue. L'absence de signe indique : crue.
0 : état stationnaire. A. signifie : année.—

Alpes

Haute-Savoie — Massif du Mont Blanc. (M. l'ingénieur Bouverot)

	1953/54	1954/55	1955/56
Argentières	—	—	—101 (4 a.)
Bois (Mer de Glace)	—	—	—182 (» »)
Bossans	—	—	231 (3 a.)
Bionnassay	—	—	— 17 (3 a.)
Taconnaz	—		311 » »
Tré la Tête	—	—	— 83 » »
Savoie — Tarentaise (M. l'Ingénieur Anchierri)			
Sources de l'Isère	—	—	—175 (6 a.)
La Vache (alias La Galise)	—	—	—
Gébroulaz	—	— 45 (2 a.)	—
Savoie — Maurienne (M. l'ingénieur Anchierri)			
Sources de l'Arc	—	—183 (5 a.)	—
Mulinet	—	— 91 » »	—
Grand Méan	—	— 25 » »	—
Evettes	—	— 28,5 » »	—
Arnès	—	— 87 » »	—
Savoie — Grandes Rousses (M. l'ingénieur Anchierri)			
Saint Sorlin d'Arves	—	— 77 » »	—
Isère — Grandes Rousses (M. l'ingénieur Huin)			
Sarennes	—X	—X	—X
Hautes-Alpes. Massif du Pelvoux (MM. Widmann et Bastide)			
Glacier Blanc	— 6	— 20	— 28
Glacier Noir	— 50	2	— 6

VARIATIONS DE LONGUEUR DES GLACIERS FRANÇAIS, EN MÈTRES.

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; A : année.—

Pyrénées, Haute-Garonne (MM. Chabrol et Sannac)			
Glaciers du Portillon	—X	—X	—X
Hautes-Pyrénées (M. l'ingénieur Chimits)			
Aussoue	0	—X	— 18 (2 a.)
Petit Vignemale	0	0	0
Oulettes de Gaube	—X	—X	— 13 (2 a.)

VARIATIONS DE LONGUEUR DES GLACIERS ITALIENS, EN MÈTRES.

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; A : année.—

Alpi Occidentali			
Alpi Marittime			
Ghiacciai	1953-54	1954-55	1955-56
Gruppo Clapier Argentera (Prof. P. Rachetto)			
Ghiacciaio del Clapier	—6	—2	—4
» » di Peirabroc	—4	—1	—2
» » di Maledia	—10	—0,5	—3
» » del Murajon	—3	—1	—2
» » di Gelas	—5	—2	—2
Alpi Graie			
Gruppo Bessanese (Mr. C. Socin)			
Ghiacciaio della Bessanese	—15	—6	—2
Gruppo Gran Paradiso (MM. A. Moretti et C. Lesca)			
Ghiacciaio del Grand Etrêt	—1	—7,5	—8
» » del Monciair	—10	—6	—18,5
» » Moncorvé	—7	—23	—10
» » del Gran Paradiso	—9	—1,5	—9
» » del Gran Neyron	—41,5	—30	—
» » del Trayo	—2	—6	—13
Gruppo Rutor (Mr. L. Peretti)			
Ghiacciaio del Rutor	—20 (?)	—10	—
Gruppo Moravidi Lechaud (Mr. C. Lesca)			
Ghiacciaio di Chavannes	—		innevato
» » del Breuil	—	—10	»
» » di Arguereuil	—	0	»

VARIATIONS DE LONGUEUR DES GLACIERS ITALIENS, EN MÈTRES.

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; A : année.—

Masiccio del Monte Bianco (Mr. C. Capello)			
Ghiacciaio di Pre de Bar	—12	0,5	—3
» » di Frebouzie	—	—X	—2 (3 a.)
» » del Triolet Alto	—0,5	—1	0
» » del Miage :			
Lingua Sinistra	—X	—15 (2 a.)	—3
» destra	0	0	—
Lago del Miage	—8	—5	—6
Ghiacciaio della Lex Blanche	—	—1	—4
Alpi Centrali			
Alpi Pennine Valpelline (Mr. I. Cossard)			
Ghiacciaio di Tza de Tzan	—22,5	—19,5	—40,5
Gruppo Cervino Testa Grigia Grandes Murailles (MM. Vanni et C. Origlia)			
Ghiacciaio di Valtournanche	3	—15	2
			—3,5 (later)
» » della Furda	0	—10	—2
» » di Tyndall	—X	—X	—X
» » di Cherillon	—1,5	1	—20
	(di potenza)	(di.potenza)	4 (later)
» » di Montabel	—X	—0,5 (2 a.)	—X
Valle d'Ayas (*Monte Rosa*) (MM. F. de Gemini et L. Valtz)			
Ghiacciaio Grande di Verra	—14,5	—15	—14
» » Piccolo di Verra	—7	—16	—3,5
Monte Rosa (MM. W. Monterin et D. de Maria)			
Ghiacciaio del Lys	—X	—30,5 (2 a.)	—11,5
» » di Indren Lingua sinistra	—0,5	—13	—11
» » Netscho		—7 (3 a.)	—2,5
» » di Piode Lingua sinsistra	innevato	3,5 (4 a.)	—47
Lingua destra	»	inenvato	—
» » del Sesia		1,5	—20,.5
» » del Belvédère	—15,5	—21	—20
			4 (later)
Alpi Retiche			
Gruppo Rambo Stella Suretta (Mr. A. Riva)			
Ghiacciaio del Ferrè	—15	—13	—3
» » di Val Loga	—14	—9	—2
» » Sud Suretta	—35	—26	—10
» » di Osrsereigls	—20 (3 a.)	—12	—3
Gruppo Disgrazia Bernina (Mr. A. Riva)			
Ghiacciaio di Ventine	—20 (4 a.)	—18	—20
» » di Caspoggio	—	—13 (4 a.)	—10
» » Fellaria W.	—14	—15	—37
» » di Scherschen	—	—195 (2 a.)	—115

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; A : année.—

Valle di Livigno (Mr. I. Bellotti)			
Ghiacciaio del Mine	—	—7	0 staz.
» » di Val Lia	—	—15,5 (2 a.)	innevato
» » di Valcardonné E.	—12	—7	—2
» » » » W.	—4	—2	—5
» » W. le Dosdé	—	—13 (2 a.)	—16
» » W. di Viola	—	—12 (3 a.)	—8
Gruppo Ortles Cevedale (Mr. R. Albertini)			
Vedretta di Sternai	—	—1 (2 a.)	—
» » di Sacrit	—	—1 »	—7,5
» » delle Marmotte	—	—1 »	—
» » di Cavajon	—	1 »	—
» » di Careser	—	—50,5 (2 a.)	—39
» » Di Lamare, Lingua N.	—	0	—
» Centr.	—	—9 (2 a.)	—35
» S.	—	1 »	—
» » Rossa	—	—5,5 »	—27
Alpi Venoste, Occidentali (Mr. A. Ricci)			
Ghiacciaio di Vallelunga	—	—37 (4 a.)	—7,5
» » di Barbadorso di dentro	—	—39 »	5
» » » » di fuori	—	—51 »	19
» » di Fontana W.	—		
» » di Giogo Alto	—	—1,5 »	—5,5
Alpi Orobie (Mr. G. Nangeroni)			
Ghiacciaio del Trobio	—	—28 »	1
» » del Gleno	—	—18 (2 a.)	0
Gruppo Adamello, versante lombardo (Mr. C. Saibaene)			
Ghiacciaio Pisgana E.	—3	—7	—?
» » » W.	—17	—13	—14
» » Salarno			—7 (3 a.)
» » Venerocolo	—10	—1,5	—1
» » Avio		innevato	innevato
Gruppo Ademallo Presanella, versante trentino (Mr. V. Marchetti)			
Vedretta Carè Alto E.	2	—0,5	—3,5
Ghiacciaio Niscli	—3	—0,5	—18
» » di Lares	—13	—5	—25
Gruppo di Brenta (MM. V. Marchetti et L. Ricci)			
Ghiacciaio Folgorida	—25	—12	—
» » della Lobbia	—29	0	1 (2 a.)
» » del Mandrone	—24	—20	—7,5
» » di Nardis	—11	—5	—X
» » d'Amola	1	—53	—17,5

VARIATIONS DE LONGUEUR DES GLACIERS ITALIENS (EN MÈTRES)

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; a : Année

» » di Cornisello	—3	—19	—2,5
» » della Presanella	—17	—6	—5
» » dei XII Apostoli	—5	—2,5	—
» » Pra Fiori	—8	1,5	—
» » di Lagol	—15,5	—15,5	—
» » Sfulmeni	—3	—0,5	0
» » di Brentei	—3	—1	—5,5
» » Tuckett	—38	—20	23
» » di Vallesinella	—	—1,5 (2 a.)	0
» » Ambiez	—	—5,5 »	—
Alpi Orientali			
Alpi Dolomitiche			
Gruppo Sorapis Cristallo (Mr. P. Nicoli)			
Ghiacciaio Sorapis Orientale	—	—0,5	0
» » » Centrale	3	0	0
» » » Occid.	—2	—0,5	—
» » del Cristallo	—	—6,5 (2 a.)	—14,5
Alpi Giulie			
Gruppo Canin (Mr. D. di Colbertaldo)			
Ghiacciaio Canni Occident.	—30	innevato	—X
» » » Orient.	—13,5	» »	—X
» » dell' Ursic	—2,5	» »	—X
» » del Montasio Occ.	—0	» »	—X
» » » » » Orient.	—	» »	—X
Appennini			
Gruppo Gran Sasso (Mr. D. Tonini)			
Ghiacciaio del Calderone	—X	—X	—X

VARIATIONS DE LONGUEUR DES GLACIERS SUISSES (EN MÈTRES)

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; A : Année

Bassin du Rhone			
Rhône	0	0	0
Mutt (Gratschlucht)	—X	—19	0
Fiesch	—2	—3,5	—3
Grosser Aletsch	—9	—44,5	—6
Lang	—5,5	—5,5	—11,5
Kaltwasser	8	0	—44,5
Allalin	—20	15	—21
Tälliboden	—10	—9	2,5
Schwarzberg (Schwarzenberg)	—8	—18	—9
Ofental	—2	—13,5	—1
Kessjen	—1	—9,5	0,5
Gorner	—29,5	—20	—24,5

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; a : Année

Z'Mutt	—6	—4	—
Findelen	—15	—9,5	—
Turtmann (ouest)	—31,5	—10,5	—29
Turtmann (Est ou Brunegg)	—53	—1,5	—93
Zinal (Durand de Tsinal)	—9,5	2	—4
Moming	—32	—7	—4
Bella Tola	—6	7	—
Moiry	—2,5	—0,5	—17
Mont Miné (anc. Ferpècle)	—18	—11	—47
Bas d'Arolla (Arolla)	—5	—34	—13
Tsidjiore-Nouve (Tsijiorenouve)	—4	—1	—6,5
Cheilon (Duran de Seillon)	—5,5	—1	—2
L'En Darrey (Lendarrey)	—7	32	—26,5
Grand Désert	—9	20	—18,5
Mont Fort	—2	3	—10,5
Otemma	—	—10,5	—20,5
Mont Durand (Val de Bagnes)	—	—12	—
Breney	—6	—9	—9
Valsorey	—29	0	—21
Corbassière	—	—3,5 (2 a.)	—18
Giètro	—	—6,5 »	—3
Saleina	—45	—19	—20
Trient	—36	—54	—20
Tsanfleuron	—34	8	—
Plan Névé (Grand)	—20	6	—
Plan Névé (Petot)	109	3,5	—
Martinets	—2	2	—
Prapio	60 (2 a.)	15	2
Sex Rouge	73 »	2,5	0
Paneyrosse	—1	0,5	—
Bassin de l'Aar			
Oberaar *	—92,5	—100	—75
Unteraar **	—16,5	—13	—1
Oberer Grindelwald	—11	—X	—X
Unterer » »	—43	—30	—15,5
Stein	—20	—17	—12
Blümlisalp	—X	0	—
Schwarz	—4	0	1
Gamchi	0	—3	—4,5
Rosenlaui	—14	—X	—
Rätzli	—	—	—68 (3 a.)
Wildhorn	—	—c	12 (3 a.)

* Retrait exagéré par l'ennoyage hydrotechnique annuel depuis 1954.
** Idem dès 1932.

VARIATIONS DE LONGUEUR DES GLACIERS SUISSES (EN MÈTRES)

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; a : Année

Bassin de la Reuss			
Griess (bei Unterschächen)	—3 (3 a.)	0	—
Wallenbühl (bei Voralp)	0,5	—	—18,5 (2 a.)
Chelen (Kehlen)	—38	4	—14
Hüfi	—8,5	—5,5	—4,5
Brunni	—14	—3	—4
Damma	—3	14	—10
St-Anna	—30	—2	—20,5
Tiefen	—22,5	10	—13
Firnälpli	—	—	21,5 (3 a.)
Griess (Engelberg)	—	—	22,5 »
Bassin de la Linth			
Sulz	—X 2(a.)	3,5	—
Bassin du Rhin			
Punteglias	—5	—7	—10
Vorab	0	—14	—19
Lavaz	—22	15	—
Porchabella	—10	—	43 (2 a.)
Verstankla	—	—X (2 a.)	—
Lenta	—9,5	—3	—8
Schwarzhorn	—6,5	—	—
Sardona	—7	—	—9 (2 a.)
Paradies	—20	—53,5	—40
Suretta	—17	23	0
Calderas	—7,5	—5	—
Piz Sol	—	—	—15,5 (3 a.)
Bassin de l'Inn			
Morteratsch	—35,5	—16,5	—20
Tschierva	—33	—37	—
Tiatscha	—17	—3	—3
Lischana	—	0 (2 a.)	—0,5
Roseg	—	—255 (5 a.)	—
Bassin de l'Adda			
Forno	—27,5	—13,5	—24,5
Palü	—542 (21 a.)	—14,5	—15,5
Paradisino	—	—	—3,5 (3 a.)
Cambrena	—	—	—2,5 »
Bassin du Tessin			
Rossboden	—4	—2,5	—
Bresciana	—6	—4	—13,5

VARIATIONS DE LONGUEUR DES GLACIERS AUTRICHIENS, EN MÈTRES.

Contrôlés par : Melle Mayer et MMrs. Schueller, Wannemacher, Prutzer, Mutschlechner, Vietoris, Paschinger, Lässer, Katschthaler, Lenzi, Pacher.—

Gletscher, Ferner, Kees	1953-54	1954-55	1955-56
Hochkönig			
Ubergossene Alm	1	2	0,5
Dachstein (Gletscher)			
Gr. Gosau	—	—2 (2 a.)	—10,5
Hallstätter	—8,5	—10	—4
Schladminger	—	—	0 (3 a.)
Silvrettagruppe (Ferner)			
Litzner NE	—11	—5,5	0
Litzner SW	—4,5	—	—
Klostertaler	—15	—	—3,5 (2 a.)
Fermunt W	—16,5	—11	—14
Fermunt E	—27	—2	—5
Bieltaler	—15,5	—1,5	—6
Jamtaler	—14,5	—3,5	—2,5
Larein	—7	—	—7,5 (2 a.)
Otztaler Alpen (Ferner)			
Weissee	—21,5	—7	—9
Gepatsch	—87	—54	—21
Hinterer Olgruben	—3	—15	0
Sexegeten	—11	—3,5	—5,5
Taschach	—68,5	—26	—23
Mittelberg	—10,5	—6	—4
Karles	—6,5	1	—1
Rettenbach	—5	0	—5,5
Hochjoch	—9	—23	0
Hintereis	—42	—15	—30
Guslar	—10	—11	—11
Vernagt	—16	—12	—2
Mitterkar	—1	—8	—2,5
Rofenkar	—5,5	—13,5	—4
Taufkar	—3	—7	—2,5
Niederjoch	—2,5	—18	—7
Marzell	—22	—	—3 (2 a.)
Schalf	—25	—	—9 »
Diem	—	—15 (2 a.)	—15,5
Spiegel	—11,5	—9	—8,5
Gurgler	—16	—14	—6,5
Langtaler	—18	—15,5	—18
Rotmoos	—48	—17	—10
Gaisberg	—16,5	—9	—10,5

VARIATIONS DE LONGUEUR DES GLACIERS AUTRICHIENS, EN MÈTRES.

Contrôlés par : Melle Mayer et MMrs. Schueller, Wannemacher, Prutzer, Mutschlechner, Vietoris, Paschinger, Lässer, Katschthaler, Lenzi, Pacher.—

Gletscher, Ferner, Kees	1953-54	1954-55	1955-56
Stubaier Alpen (Ferner)			
Sulztaler	—46 (2 a.)	—	—43,5 (2 a.)
Bockkogel	—32,5 »	—	—1,5 »
Schwarzenberg	—10 »	—	—12 »
Bachfallen	—1,5 »	—	—14,5 »
Längentaler	—1 »	—	2 »
Lisenser	—20,5 »	—	—24,5 »
Alpeiner	—25,5 »	—	—29,5 »
Berglas	—34 »	—	—170 »
Hochmoos	—2	—	—
Daunkogel	—11,5	—11	—9,5
Schaufel	6	—6,5	—12,5
Fernau	—4	—6,5	—4.5
Sulzenau	—10,5	—24,5	—36
Grünau	—6,5	—11	—12,5
Grübl W	—7	—	—
Grübl E	0	—	—
Simminger	—7,5	—	—2,5 (2 a.)
Zillertaler Alpen (Kees)			
Waxegg	—10	—2	—4
Horn	—39	—30	—23,5
Schwarzenstein	—6,5	—9	—5
Venedigergruppe (Kees)			
Krimmler	—11 (2 a.)	—	—3,5 (2 a.)
Obersulzbach	—52,5	—	—52 »
Untersulzbach	—77,5 (2 a.)	—	—31,5 (2 a.)
Habach	—19 »	—	—6,5 »
Umbal	—15,5 »	—	—
Simony	—16,5 »	—	—
Maurer	—66,5 »	—	—
Dorfer	—35,5 »	—	—
Mullwitz (Zettalunitz)	—52,5 »	—	—
Frosnitz	—19 »	—	—
Schlaten	—15,5 »	—	—36,5 (2 a.)
Viltragen	—29,5	—	—23,5 »
Glocknergruppe (Kees)			
Karlinger	—23,5 (2 a.)	—	—20,5 (2 a.)
Bärenkopf	—13 »	—	—14 »
Klockerin	—12,5 —	—	—13 »
Pasterzen	—11	—7,5	—12.5

VARIATIONS DE LONGUEUR DES GLACIERS AUTRICHIENS, EN MÈTRES.

Contrôlés par : Melle Mayer et MMrs. Schueller, Wannemacher, Prutzer, Mutschlechner, Vietoris, Paschinger, Lässer, Katschthaler, Lenzi, Pacher.—

	1953-54	1954-55	1955-56
Wasserfall	—	—	—2,5 (3 a.)
Freiwand	—	—	—1 »
Pfandlscharten	—6,5	—	0 (2 a.)
Ankogelgruppe (Kees)			
Grosselend	—5,5	—12,5	—8
Kleinelend	—	—	—1,5 (4 a.)
Kälberspitz	—13	—2,5	—9,5
Tripp W	—	—0,5 (3 a.)	—2,5
Hochalm	—2	—1 (?)	—1
Winkel	—10,5	—11	—13

VARIATIONS DE LONGUEUR DES GLACIERS NORVÉGIENS

	1953-53	1954-55	1955-56
Jostedalsbreen			
Böyumbreen	—12	7	7
Store Suphellebre	—10	—3	—7
Austerdalsbreen	—60	—43	—43
Tunsbergdalsbreen	—21	—12	—10
Nigardsbreen	—41	—72	—54
Fabergstölsbreen	—40	—62	—34
Lodalsbreen	—30	—54	—52
Stegaholtbreen	—27	—31	—29
Briksdalsbreen	—13	—8	7
Jotunheimen			
Styggbreen	—20 (2 a.)	—18	—11
Tverrabreen	—16	—12	—13
Veslejuvbreen	—13 »	3	—4
Storjuvbreen	—41	—31	—33
Nordre Illabre	—11	—6	—18
Söndre Illabre	—13	—5	0
Storbreen (Leirdalen)	—18	—13	—15
Leirtjönnbreen	—10	—	—27 (2 a.)
Böverbreen	—9	—	—33 »
Slettmarkbreen	—20 (2 a.)	—	—18 »
Langedalsbreen	—29 »	—	—18 »
Styggedalsbreen	—14	—6	—9
Svartisen			
Osterdalssisen	—32	—51	—12

VARIATIONS DE LONGUEUR DES GLACIERS SUEDOIS

	1953-54	1954-55	1955-56
Stor glaciären	—16	—9	—7

VARIATIONS DE LONGUEURS DES GLACIERS ISLANDAIS (EN MÈTRES)

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; a : année

	1953-54	1954-55	1955-56
Drangajökull			
Kaldalonsjökull	—3	—12	—75
Leirufjardarjökull	—68 (2 a.)	—110	—30
Reykjarfjardarjökull	—228	—72	—26
Snoefellsjökull			
Hyrningsjökull	—30 (2 a.)	—	—65 (2 a.)
Jökulhals	—260 »	—	—10 »
Blagilsjökull	—80 »	—	—
Eyjafjallajökull			
Gigjökull	313 (8 a.)	—	—
Myrdalsjökull			
Solheimajökull W.	—48	—26	—14
» » E.	—10	—20	—20
Jökulhöfuo	—21	0	0
Vatnajökull			
Skeidarárjökull W.	0	—40	—120
» » » E.	—29	—93	—58
Morsárjökull	—35	—88	—102
Skaftafellsjökull	—49	—31	—47
Svinafellsjökull N.	3	—15	—1
» » »S.	7	3	4
Virkisjökull (Falljökull)	10	5	0
Kviarjökull	—23	—32	—20
Hrútárjökull	—22	—34	—3
Fjallsjökull	—14	—4	0
Breidamerkurjökull W.	7	—28	—19
» » » E.	133	—49	—20
Fellsjökull	—	—	—80 (4 a.)
Brókarjökull	—15	33	37
Birnujökull	—5	—2	—1

VARIATIONS DE LONGUEURS DES GLACIERS ISLANDAIS (EN MÈTRES)

Le signe — signifie : décrue; l'absence de signe : crue; 0 : stationnaire; a : année

	1953-54	1954-55	1955-56
Heinabergsjöklar			
Eyvindstungnakollur (Heinabersgjökull)	—8	—10	—2
Syori jölkul (Skálafellsjs.)	—34	—25	—12
Nyrori jökull (Heinabergsj.)	—20	—35	—20
Fláajökull W.	—41	—47	—7
» » E.	—41	—	—
Hoffellsjökull W.	—17	—46	—62
» » E.	—13	—4	—10
Hofsjökull			
Nauthagajökull	—6	—4	—40
Mulajökull	160	15	—44
Langjökull			
Hagavatnsjökull syori (S-glacier)	—	—	—168 (6 a.)
Fulakvislarjökull (Pjófadölum)	—	—	—163 (5 a.)
Kerlingarfjöll			
Lodmundarjökull	—	—	—5 (5 a.)
Nordurland			
Gljúfrarjökull	—	—68 (2 a.)	—25

VARIATIONS DE LONGUEURS DES GLACIERS D'EUROPE

Régions	Nombre de glaciers observés	Récapitulation en crue	stationnaires	en décrue
France				
1953/54	7	—	2	5
1954/55	14	1	1	12
1955/56	14	2	1	11
Suisse				
1953/54	73	5	3	65
1954/55	76	19	7	50
1955/56	67	8	4	55
Italie				
1953/54	62	4	3	55
1954/55	90	6	5	79
1955/56	83	7	7	69

Autriche				
1953/54	74	2	1	71
1954/55	44	2	1	41
1955/56	71	2	5	64
Norvège				
1953/54	22	–	—	22
1954/55	18	2	—	16
1955/56	22	2	—	20
Islande				
1953/54	34	7	2	25
1954/55	25	4	1	20
1955/56	35	2	3	30
Suède				
1953/54	1	—	—	1
1954/55	1	—	—	1
1955/56	1	—	—	1
TOTAUX				
1953/54	273	18	11	244
1954/55	268	34	15	219
1955/56	293	23	20	250

GLACIATION OF THE TORNGAT MOUNTAINS, NORTHERN LABRADOR [1]

J. D. IVES [2]

Abstract

The field work was carried out during the summer of 1956, and the paper is based upon this work together with a preliminary study of the aerial photographs.

The highest summits of the Torngat Mountains were completely submerged at the Wisconsin Maximum by ice from the west moving toward the Atlantic Ocean. Erratics found at heights of up to 5.000 feet imply that the thickness of the ice above the summits was considerable.

Following this stage it is envisaged that for a long period the level of the continental ice stood at the 2.000-2.200 feet level, and thereafter the ice melted predominantly by downwasting in situ when the snow line passed above the land surface. Detailed study of striations, erratics, and associated glacial features revealed that the final movement of ice was from the west through the mountains, and that some flow continued in the through-troughs after the local valleys had almost completely emerged. Thus ice lakes were damned in the tributary valleys laterally against continental ice in the through-troughs, and no local resurgence of glaciers occured after the dissipation of the inland ice.

It is concluded that the Torngat Mountains were not a local centre of glacial outflow in late-Wisconsin times, with the exception of small cirque glaciers. Moreover, the hypothesis is presented that with a general lowering of the present snow line of Labrador-Ungava instantaneous glacierization across large areas of the plateau to the south-west would occur before the growth of any significant glaciers in the Torngat Mountains.

A preliminary study of the semi-permanent snow beds and the present snow line in Labrador-Ungava tends to support this hypothesis. Further study, however, is necessary before a conclusive statement of changing Wisconsin conditions can be presented.

This paper was read in the absence of the Author by Dr. Svenn Orvig.

Introduction

The writer, (accompanied by his wife, Pauline Ives, as field assistant), worked in the Torngat Mountains of Northern Labrador between 27th July and 16th September 1956. The work was made possible by research grants from the Banting Fund, administered by the Arctic Institute of North America, and from the McGill-Carnegie Arctic Research Programme. Transport from Goose Bay to and from the Torngats was generously provided by the British Newfoundland Corporation.

Upper Kangalaksiorvik Lake, some 15 miles west of Seven Islands Bay in latitude 59°22′ north, was selected for the site of a base camp, and this was established with the aid of a Beaver aircraft on 27th July. In addition, the Beaver was able to lay two food and fuel caches, one at the northeast end of Lower Komaktorvik Lake, and the other by the shore of a small lake on the Quebec side of the boundary, some 15 miles west of base camp. From these three centres an area of about 600 square miles, principally within the «Central Range» of the Torngat, was reconnoitred on foot.

Despite the short and relatively stormy season, it proved possible to complete a large part of the original programme, although an investigation of the cirque glaciers was curtailed by the persistence throughout the summer of a considerable thickness of the 1955-56 snow cover on their surface. After 20th August this was augmented by frequent falls of fresh snow.

[1] Previously published in 'Arctic, Vol. 10, pp. 76-87.
[2] Mc Gill Sub. Arctic Research Laboratory, Scheffcrville P. Q. Canada.

The following information is of a preliminary and reconnaissance nature, although it may be anticipated that future work will not greatly affect the general conclusions. The paper is based primarily upon a rapid review of the field notes and a cursory examination of the vertical air photographs from Saglek Bay northwards to Cape Chidley. It is hoped that this work will be intensified and extended in the future. The writer wishes to acknowledge the Report of Mr. Murray Piloski of the British Newfoundland Corporation for information on the bedrock geology of the area.

Objectives of the Field Work

Two generally accepted, and yet untested, theories, which have played a fundamental part in the modern concept of the glaciations of Labrador-Ungava, and of north-eastern North America as a whole, prompted the field investigations outlined below.

In 1933, Odell, refuting the earlier work of Daly (1902), Bell (1882-84) and Coleman (1921), stated that the entire area of the Torngat Mountains had been completely inundated by continental ice at the Wisconsin maximum. Despite the slender evidence which Odell put forward, his concept, perpetuated by the writings of Tanner (1944) and Flint (1943 and 1947) has gained almost universal acceptance. Odell's main criticism of Coleman is based upon the assumption that the felsenmeere so widely distributed above the 2,000 — 2,200 foot level was formed by frost action in post-glacial times.

Flint (1943, 1947, 1952 and 1953) has contributed extensively to the present understanding of the growth and disappearance of the Quaternary ice sheets. In considering the general climatic deterioration at the onset of glacial times, Flint envisaged the intial growth of vigorous glaciers in the coastal mountains of north-eastern North America stretching from Labrador to Ellesmere Island. Flint states that the glaciers forming on these mountains would flow down their western flanks to accumulate as piedmont lobes and ultimately build up into an ice cap of continental proportions. During this growth the ice divide would move westwards from the mountains resulting in a reversal in the direction of flow across the mountains towards the east.

Flint pictured the reversal of these conditions towards the close of each glacial epoch, with the coastal mountains serving as the final centres of glacial outflow. This thesis apparently considers conditions in northeastern North America as the mirror image of those in northwestern Europe where the theory is based upon solid field evidence. Flint believed that the area west of the Torngat would prove of critical importance in the testing of this theory in the field, a belief which greatly influenced the general localisation of this study.

The main objective of the present work, therefore, was to attempt to establish the role of the Torngat in the glacierization of Labrador-Ungava, and in particular to assess the course of events in late-Wisconsin times with respect to the final disappearance of both continental and local ice masses.

Geological Background

North of Nachvak Fiord the Labrador peninsula is composed of metamorphic and igneous rocks of Precambrian age. Between Nachvak Fiord and the Kangalaksiorvik lakes the main structural trend is north-northwest to south-southeast roughly parallel with the coastline, and metamorphic zones of varying degree of intensity follow this trend. Much of the bedrock is composed of various types of gneiss, a finely-banded garnetiferous gneiss and a granite-gneiss being particularly abundant.

Fig. 1 — Glacial striations and grooves in banded gneiss. The Surface is relatively unweathered as it lies beneath high lake level. The glacial markings cross the lineation at right angles. 2 Aug. 1956

Fig. 2 — Roches moutonnées and perched blocks in the Abloviak-Kangalaksiorvik trough looking east. Kangalaksiorvik River and terraces in middle distance. 4 Aug. 1956

The rocks are predominantly coarse-grained and weather into a very rough surface upon which glacial markings are seldom clearly preserved, even on the lower ground which was cleared of ice most recently (Fig. 1). The entire area is cut by two series of basic dykes.

In the vicinity of the Kangalaksiorvik lakes the dominant structural trend swings round in a great arc and becomes roughly east-west on the Quebec side of the peninsula. Piloski has described this as a possible drag fold and it has an important influence upon the present relief. A major trough-like valley, passing inland from the head of Kangalaksiorvik Fiord, closely follows this trend cutting through the height of land and passing westwards into Abloviak Fiord. Its highest point lies below 1,000 feet, although the surrounding mountains exceed 4,000 feet.

The major faults follow the dominant structural trend, although a second trend, roughly at right angles to the first, is readily apparent in the topography.

Physiographic Evolution

Until a more detailed examination of the available material has been made the outline of the evolution of the Torngat Mountains given by Cooke (1930)and supported and extended by Odell (1933) and Tanner (1944), will be adopted as the most satisfactory. Very briefly this envisages the predominance of erosional processes over sedimentation since Precambrian times. In late-Tertiary times the Labrador peneplain was uplifted to a considerable height and tilted so that a high escarpment bordered the Atlantic Ocean and the plateau surface sloped down gradually towards the west. This event has severely conditioned the subsequent evolution of the area, and in this connection it is emphasized that the Torngat Mountains are not a mountain range in the true sense of the word, and in particular that there is no western flank, but merely a gentle plateau slope which passes gradually beneath the water of Ungava Bay. The terminology of «Coastal Range» and «Central Range» proposed by Odell (1933) is therefore misleading and extremely unfortunate. Throughout this report, however, Odell's terminology is retained for the purpose of clarity when referring to different localities.

It is assumed that prior to the onset of glacial times the uplifted peneplain had been subjected to the processes of a fluvial cycle of erosion for a sufficient length of time to allow a late youth to early mature stage of dissection to develop. This created a very rugged mountain country bordering the Atlantic Ocean, grading westwards into a slightly dissected plateau.

This is an over simplification of the true pattern of development, but little more need be added for the present purpose other than emphasizing that the late-Tertiary uplift was by no means uniform, and that warping and faulting occurred, the results of which are clearly visible in the landscape today.

The precise number of glacial periods which have affected the area is unknown, and so far no stratigraphic proof of more than one glacial period has been found in Labrador-Ungava. It is assumed, however, that the sequence of events in the Torngat closely paralleled the developments which are comparatively well known in southern Canada and northeastern United States. Morphological evidence, in the form of valley-in-valley cross sections and the floors of glacial valleys trenched by deep waterworn gorges of presumed pre-Wisconsin age, was found in the Torngat to support this assumption. The evidence indicates that the present landscape owes its appearance to two and possibly three glacial and inter-glacial cycles.

The actual amount of erosion attributable to glacial action is a matter of consi, derable controversy, especially when viewed in the light of recent work (Battle 1952; McCall, 1952; Boyé, 1950; Cailleux, 1952; etc.) much of which tends more

Fig. 3 — Boulder fields of glacial moraine from which the fines have been washed by melt-water. Altitude about 1,500 feet. 2 Aug. 1956

Fig. 4 — Frost-riven bedrock (felsenmeere) on summit of peak in «Central» Torngat Mountains. Altitude about 4,000 feet. 2 Aug. 1956

and more to question the effectiveness of glaciers as agents of erosion. It is assumed that little or no erosion of the open plateau surface was accomplished by the passage of the continental ice. Thus Odell's designation of «glacial peneplain» (1933, p. 209) is not accepted. A considerable amount of vertical and lateral erosion was probably accomplished in the great through-troughs and valleys, where the ice was thick and the speed of flow accentuated by the topographical restrictions.

Fluvial erosion in pre-glacial and inter-glacial times, together with frost action and mass movement, and sub-glacial fluvial erosion, has probaly produced more important results by volume of material removed than have the glaciers. The presence of the glaciers, apart from their active participation, has been necessary for the production of the characteristic «glaciated» appearance of the Torngat Mountains today. It is not enough to state that the glaciers are ineffective as agents of erosion; an alternative method must be proposed which will adequately account for the U-shaped cross section with truncated spurs, the over-deepening and production of rock basins and valley steps, and the numerous other features so closely associated with «glaciated» mountains. Undoubtedly much more research is needed into the mechanisms of glacier flow and glacial erosion, but until conclusive evidence is forthcoming to the contrary the critical acceptance of the theory of glacial erosion seems advisable.

The Extent of the Wisconsin Glaciation

The discussion is restricted to a consideration of the Wisconsin Glaciation, first, because the pre-Wisconsin evidence is so far indistinguishable, and secondly, because the Wisconsin was probably the least severe of all the glacial periods so that the conclusions reached here can be regarded as minimal for the greatest extent of continental ice in Quaternary times.

Coleman (1921), and all workers before him (Daly, 1902; Bell, 1882-84) believed that the Torngat maintained a local centre of glaciation throughout Wisconsin times, that the penetration of continental ice from the west was limited, and that the higher summits, above about 2500 feet above present sea level, remained as nunataks even at the height of the ice flood.

Coleman's argument is based largely upon the sharp mountainous forms of the upper summits, which he believed would not have withstood inundation by continental ice, and upon the presence, on the higher surfaces, of a heavy mantle of frost-riven bedrock, or felsemeer, which he considered to be of pre-Wisconsin age. Odell (1933), on the other hand, argued that the felsenmeer was formed in post-glacial times by vigorous frost shattering, which thus invalidates Coleman's conclusion, and, despite the intensity of frost action, Odell found evidence of glacial polishing and grooving at an altitude of 4,700 feet.

Apart from the theoretical challenge by Dahl (1946 and 1947) who suggests that the coastal summits above 3,000 feet must have remained as nunataks upon a consideration of the marginal surface slope of inland ice sheets, the conclusions of Odell have remained largely undisputed (see, however, Mercer, 1956). Work in Iceland, where it was possible to prove conclusively that the high coastal mountains remained as nunataks throughout the glacial epoch, induced the writer to consider carefully the objections of Dahl, and to attempt to reassess the field evidence before making a final conclusion.

Thirteen of the major summits in the «central» Torngat, between Chasm Lake and Upper Kangalaksiorvik Lake, were ascended and their surfaces were subjected to minute examination. In addition innumerable spot examinations were made for evidence of glacial action throughout the area studied. From this two important generalisations can be made. First, in the eastern sector, between Mount Tetragona

Fig. 5 — Perched blocks on glaciated summit of ridge above the confluence of Abloviak River with south bank tributary. Note the absence of frost-riven blocks. Altitude about 2,900 feet. 7 Aug. 1956

Fig. 6 — Felsenmeere on summit $^1/_2$ mile west of summit in Fig. 6. Occasional large boulders possibly emplaced from ice cover. «Ungava Bay Plateau» in background. Altitude about 2,900 feet. 7 Aug. 1956

and the Quebec boundary, there is a marked contrast between surfaces above and below the 2,000—2,200 foot level. Below this level evidence of widespread glaciation, as noted by Coleman (1921), was abundant, mainly in the form of erratic and perched blocks, glacial striations (Fig. 1), ablation moraine (Fig. 3), roches moutonnées, and fluvioglacial deposits. Above this level evidence of glaciation was almost entirely lacking. Secondly, on the western side of the Labrador-Quebec boundary where the mountains merge imperceptibly with the «Ungava Bay Plateau» conclusive evidence of glaciation was found up to a height of 3,000 feet above present sea level. Superficially it appeared that Coleman's interpretation, with this modification, was generally correct. Nor would it be anticipated that Odell's poorly preserved striations could have survived the vigorous frost action which he himself invokes to explain the formation of the felsenmeer in post-glacial times. This is especially emphasized when it is pointed out that the bedrock upon which Odell found indications of glacial striations was a finely-banded coarse-grained gneiss which weathers to produce coarse, north-northwest to south-southeast ridges and furrows, and a similar set at right angles along the cross jointing, all of which closely resemble poorly preserved striations, and upon which it is impossible to distinguish glacial markings when in a weathered state.

Before proceeding further it is necessary to consider the weight of the evidence which is placed upon the presence of felsenmeer upon the upper surfaces. Examination of the superficial cover throughout the Torngat revealed that, excluding obvious till and fluvio-glacial deposits, there are two distinct types of boulder fields (felsenmeer). Below the 2,000 — 2,200 foot level the mass of boulders is characterised by a sub-angular to sub rounded form (Fig. 3), although occasional boulders have been split subsequent to the rounding process. Above this altitude and away from talus slopes the surface cover of rock debris is distinctly angular (Fig. 4). The latter is described as true felsenmeer, the product of frost-shattering on bedrock; the former is believed to be glacial moraine from which all the fine material has been washed away by running water.

A further examination of the felsenmeer takes the argument to its conclusion. The summit ridge overlooking the junction of the Abloviak River with its main south bank tributary has two summit areas separated by a slight saddle, each about 2,800 feet above sea level. The ridge follows the east-west structural trend so that the composition of the bedrock of which the two summits are composed is identical — a coarse-grained gneiss. The eastern summit is a ridge of bedrock which is littered with perched blocks which have undoubtedly been deposited by continental ice (Fig. 5). In contrast (Fig. 6) the western summit is broad and level, about 600 yards across, and is covered with a deep mantle of frost-riven bedrock. Erratics were found amongst the angular blocks. If the felsenmeer is truly of post-glacial age it is difficult to account for the relatively unweathered appearance of the eastern summit (Fig. 5). That this summit has been swept clear of superficial material, implies that, as the rock type is the same, its original cover of felsenmeer has been formed in pre-Wisconsin times and that it has been removed subsequently by the same agency which deposited the erratic blocks. The broader summit (Fig. 6), presumably because of a difference in local conditions, of which the areal extent is probably important, retained all or most of its cover of felsenmeer, which, however, may have been slightly reworked by the passage of ice.

The evidence already presented appears to invalidate the conclusions of both Coleman (1921) and Odell (1933), and the problem would have remained unsolved but for the eventual discovery of unequivocal erratic bocks (Fig. 7) on two of the major summits ascended, one at a little over 4,000 feet, and the other at the 5,000 foot level.

The general scarcity of erratic material on the higher summits must not be

Fig. 7 — Erratic block of granité-gneiss on summit in «Central» Torngat Mountains. Altitude about 4,000 feet. 27 Aug. 1956

Fig. 8 — Dissected perched delta and glacial lake shoreline in south bank tributary valley of the main Abloviak-Kangalaksiofvik trough. 10 Aug. 1956

over-emphasized as, from a consideration of the morphology, the emplacement of erratics by continental ice flowing from lower land west of the Torngat would be a matter of some difficulty. Any erratic material found at over 4,000 feet must have been dragged up in the bottom layers of the continental ice. This movement would need a vigorous flow of ice across the summits, which in turn implies that the summits in question were submerged beneath a considerable thickness of ice.

The morphology of the two mountains upon the summits of which the erratic blocks were found is interesting in this connection. Each summit was broad and level with a gently sloping west or southwest ridge. On one, which lies immediately north of Precipice Mountain in the Komaktorvik lakes area, the southwest ridge is broad and slopes gently from the summit, at a little over 4,000 feet, down to 2,700 feet. A trail of erratics was found from the 2,700 foot level all the way to the summit. The second summit is similar in form; it exceeds 5,000 feet, and lies immediately north of Chasm Lake. It is considered that such gentle slopes in the direction from which the ice came would be particularly favourable to the emplacement of erratic blocks, whereas it seems likely that steep onset faces would be inimical to this process.

The erratic material was of granite-gneiss composition and was found to be resting upon finely-banded garnetiferous gneiss. As the bedrock geology west of the Labrador-Quebec boundary is little known, this material may only have travelled a short distance, especially as granite-gneiss is a very common rock in the Torngat area. However, its presence was sufficient to indicate that continental ice had passed over mountains exceeding 5,000 feet and it is considered that the thickness of ice above the summit must have been in the order of 1,000 feet in order to promote the vigorous movement necessary for the emplacement of the erratics. The erratics were relatively unweathered compared with the bedrock and felsenmeer upon which they were resting, and it is assumed that they were deposited in post-Sangamon times.

From the evidence outlined above the following conclusions are put forward:—

1) The formation of felsenmeer pre-dates the Wisconsin Glaciation.

2) Boulder fields of sub-angular to sub-rounded form below 2,000 to 2,200 feet are washed glacial moraine.

3) During the maximum of the Wisconsin, and hence during each preceding glacial period the highest summits of the «central» Torngat were submerged beneath continental ice about 1,000 feet in thickness.

4) The maximum stage of inundation was relatively short-lived, and following this stage the higher summits stood as nunataks above the ice sheet for a considerable period.

5) A position of equilibrium was maintained for a long period when continental ice, flowing eastwards through the great west-east troughs and diffluence passes stood at the 2,000 — 2,200 foot level in the east, and up to 3,000 feet on the western slope of the «Ungava Bay Plateau» where the damming effect of the higher land caused the eastward moving ice to be piled up to a greater height. During this period the local cirque glaciers would be very active and the sharp serrated ridges and peaks would receive their final etching.

The Extent of the Wisconsin Glaciation on the Coastal Summits

So far the discussion has not included the so-called «coastal range» of Odell (1933) where summits of the Four Peaks group, Mount Tetragona, and southwards, exceed 3,500 to 4,500 feet in close proximity to the present coastline. As no definite evidence of the glaciation of these summits was obtained the discussion must follow the theoretical lines of Dahl (1946 and 1947).

Dahl objected to the conclusions of Odell (1933) and Tanner (1944) on the thickness of the Wisconsin ice sheet on theoretical grounds based upon the measurement of the marginal slopes of present and past ice sheets. Dahl suggests that the maximum slope of an ice sheet bordering a deep ocean would be 1 : 100, and upon this basis sharp coastal summits exceeding 1,000 meters within 100 kilometres of the margin of such an ice cap would remain as nunataks. Assuming the presence of a floating ice shelf, Dahl allows that the coastal summits could slightly exceed 1,000 meters and still fail to project as nunataks. Mercer (1956) has reiterated this argument and applied it to the southeast peninsula of Baffin Island as well as to the Torngat Mountains.

The first edition of the Canadian Hydrographic Charts, Nos. 4775 and 4776, published in 1955, clearly indicates that a broad continental shelf fringes the Northern Labrador coast, exceeding 100 miles in width in places. Eastward of the Torngat the shelf extends for 90 to 110 miles reaching a depth of 140 — 150 fathoms along its outer margin before dropping off steeply into very deep water. It is contended that the continental shelf would allow the accumulation of an enormous mass of ice, and if Dahl's marginal slope of 1 : 100 is applied to a continental ice sheet, the outer margin of which floats on the deep water bordering the shelf, such coastal summits as Mount Razorback would certainly be submerged, and it is most probable that the higher summits, slightly farther inland would also be submerged, if only by a relatively thin cover of ice.

It is seen, therefore, that the evidence which supports the conclusion that the interior summits were covered by ice about 1,000 feet thick at the Wisconsin Maximum, suggests that the extreme coastal mountains were also submerged.

Directions of Ice Movements in the Torngat

The main movement of ice through and over the Torngat Mountains was from a westerly direction. As the bottom layers of the ice were strongly controlled by the topography, and as no striations were found on the summits examined, it has not been possible to give a more precise direction of movement than this. Although the trends of the striations in the valleys are, therefore, of limited value, they are instructive as, collectively with directional evidence gained from the examination of roches moutonnées, erratics and fluvio-glacial deposits, they provide a picture of conditions during the latter half of Wisconsin times.

As suggested above, it is envisaged that for a considerable period following the maximum extent of the Wisconsin ice sheet the higher summits projected as nunataks. During this time the pattern of movement clearly suggests that the Torngat were partially inundated by a great anastomosing system of glaciers flowing from an area to the west through every available trough, valley, pass and col. towards the Atlantic Ocean. This was slightly augmented by local ice supplied primarily from the numerous cirques.

Several interesting cirques were examined and these showed every stage in the transition from a self-contained cirque with the backwall unbroken to an ice diffluence valley resulting from the progressive breaching of the backwall by a combination of cirque erosion and penetration by ice moving from the west.

It is emphasized that movement of ice from the west predominated except in the limited cases of local movement within the cirques. In no instance was there found any evidence of the reversal of flow towards the west from a theoretical local ice cap, and in only minor instances does the morphology of the western part of the Torngat coincide with westerly drainage. The three east bank tributaries of the main Abloviak tributary valley are broad and typical U-shaped valleys which trend

roughly east-southeast to west-northwest. The two largest valleys enter the main valley at grade, the third hangs some 800 feet above the floor of the main valley. From a consideration of their morphology at some period they contained westerly flowing glaciers.

No moraines nor striations were found in these valleys, but even allowing the morphology of the valleys as evidence of local glacierization, and assuming that these valleys contained glaciers in late-Wisconsin times, this is of very limited extent and does not invalidate the general conclusion that no major reversal of flow occurred toward the close of Wisconsin times.

Examination of the movement of erratics was made in detail in certain limited areas. For example, all anorthosite erratics were found eastward of their parent outcrops. The same conclusion was drawn from an examination of the eastern margin of the north-south zone of garnetiferous gneiss which extends from Nachvak Fiord to the Kangalaksiorvik lakes. Lack of clear differentiation between the various rock types of the metamorphic zones did not allow further expansion of this method in the time available. Occasional blocks of crystalline quartzite were found in the Kangalaksiovik-Abloviak trough and below the Komaktorvik lakes, which may have been derived from the outcrop of crystalline quartzite shown on Piloski's map northwest of the head of Upper Komaktorvik Lake. If this is so it suggests a circuit movement of the bottom ice conditioned by the major topographic forms. Blocks of a pudding stone conglomerate were found along the floor of the main south-bank tributary of the Abloviak River. This rock was not found in situ and it is possible that it may represent blocks of indurated moraine.

Late-Wisconsin Conditions

The predominance of the west to east direction of ice movement is significant in a consideration of the process of deglacierization of the Torngat. A series of critical localities, such as the thresholds of north-facing cirques tributary to major west-east valleys, the confluence of tributary valleys and the main through-troughs, and several of the main ice diffluence cols, was selected and studied in an attempt to assess the extent of any late-Wisconsin resurgence of local movement once the main ice streams from the west had disappeared. All the evidence suggests that such a resurgence did not occur, and that the final movement of ice was from the west. This was illustrated by an examination of the threshold of the cirque on the north face of Mount Tetragona which contains Bryant's Glacier. The present snout stands at about 1,400 feet above sea level and below it three morainic arcs extend northwards; the terminal moraine, lying at about 1,100 feet above sea level, is situated almost one mile north of the snout. This moraine probably marks the historical maximum of the glacier, although it may possibly represent a late-Wisconsin stage.

The cirque opens out into a major U-shaped valley which cuts from southwest to northeast through the Tetragona mountain group. Clear evidence was found in this valley, in the form of erratics, roches moutonnées and striations, that the final movement of ice was towards the northeast, and striations were found within 100 yards of the terminal moraine of Bryant's Glacier. Thus it is apparent that in late-Wisconsin times the cirque glacier was not sufficiently vigorous to extend beyond the limit of its own cirque. Similar evidence was found throughout the area and indicates the insignificance of a late-Wisconsin outflow of ice from the Torngat.

This was further substantiated by the presence of lake shorelines in valleys tributary to the main through-troughs, indicating that ice-dammed lakes existed in the local valleys while continental ice still remained in the main valleys. A well formed shoreline was observed in the south bank-tributary of the Abloviak Valley extending

six miles upstream at an altitude of almost 600 feet above sea level. A series of perched deltas, subsequently dissected, were found on each side of the valley and coincided with the shoreline (Fig 8). Subsequent lowering of the trunk glacier in the Abloviak Valley appears to have been gradual as no lower shorelines nor perched deltas were found. It is clear, however, that in the main valley, once the divide area had emerged from the ice, a second lake was formed to the west of the divide which spilled over eastwards into the Kangalaksiorvik River.

Examination of the aerial photographs reveals the presence of lake shorelines in other parts of the Torngat. Mr. Piloski and Dr. Wheeler have informed the writer that similar ice-dammed lakes were formed in the Saglek Fiord-Korok River area farther south.

For the present no definite chronological correlation between the lake shorelines, overflow channels and marginal drainage channels can be attempted. However, the main lines of the deglacierization can be envisaged.

Following the emergence of the summits as nunataks the Torngat remained in a stage of partial submergence for a considerable period. With the general climatic amelioration which heralded the close of Wisconsin times the snowline in Labrador-Ungava gradually became higher. A critical height would be reached after which a relatively slight vertical movement would lift the snowline above a vast area of the ice cap, and in a very short time the entire cap would become climatically dead. Thus the early stages of thinning would be gradual but the final stage would result in rapid thinning of the glaciers and the ice sheet itself. At this stage it is believed that the main trunk glaciers were still of the order of 2,000 feet in thickness and continued their flow towards the east for some time. Progressively the supply of ice would diminish and they would eventually melt in situ, the floors of the deepest valleys being the last areas to be free of ice.

This condition is clearly seen in the Komaktorvik-Chasm lakes area where sub-parallel marginal drainage channels sloping gently down valley can be traced from an altitude of about 2,000 feet down to the valley floors. The average vertical interval between the channels is about 15 feet, and if, as maintained by Mannerfelt, they are annual features, the final wastage of the 2,000 feet thick Komaktorvik and Chasm glaciers appears to have been completed in less than 150 years, although this does not allow for temporary climatic fluctuations.

This phenomonal rate of disappearance of ice and the production of huge volumes of melt-water is in accordance with the evidence that below the 2,000 to 2,200 foot level the land has been thoroughly washed by strong melt-water torrents.

Typical dead-ice topography in the floors of the main valleys indicates the location of the final ice remnants — at those places where the ice reached a maximum thickness in Wisconsin times and where the supply from the west was probably the most vigorous.

Finally, it is probable that when the deglacierization was complete the cirque glaciers also melted away, and that the present diminuitive cirque glaciers have developed since the post-glacial Thermal Maximum, that they reached their maximum historic extension between 1600 and 1850 A.D., during the so-called «Little Ice Age», and at present are possibly just emerging from a period of rapid recession.

Following the emergence of the land from the Wisconsin ice masses a marine transgression occurred, reaching a minimum height of 205 to 225 feet above present sea level. The assumption that the original depression, due to the thickness of the ice load, and hence the subsequent recovery, was greatest in the west is supported by scattered pieces of field evidence. Old river terraces in the Abloviak Valley were seen to be inclined gently towards the west, that is, in a down-valley direction. Mr Piloski has also described a lake shoreline above Nakvak Brook, which drains into the northern arm of Saglek Bay, to be inclined towards the west.

The Role of the Torngat as a Possible Centrale of Ice Dispersal

From the evidence presented above it appears that in late-Wisconsin times no centre of ice dispersal existed in the Torngat Mountains. Before the possible role of the Torngat in the initiation of the Wisconsin ice sheets can be considered the physical character of the area must be re-emphasized.

Flint (1943) tends to regard the Torngat as a mountai range, and if this were the case then his theory of the initial accumulation and dispersal from such a range is logical. As has been stated above however, Odell's terminology of «Coastal» and «Central» *ranges* is especially unfortunate as the Torngat are merely the dissected eastern edge of an uplifted and tilted peneplain. Above all, the western slope is extremely gentle and any such description as «western flanks» is erroneous. Secondly, a vast area of the Labrador-Ungava plateau exceeds 2,500 feet in altitude, which, when considering the areal extent and position in relation to the source of precipitation, is much more significant as an initial accumulation area than are the Torngat.

Preliminary investigations of the snowline in Labrador-Ungava, after the style outlined by Professor Gordon Manley in England (1949), also lend support to this concept. Semi-permanent snow beds in the Knob Lake area at an altitude of 2,000 feet above sea level allow a rough estimate to be made of the present height of the snowline above the land surface. This is approximately 2,000 feet or 5,000 feet above sea level. In the Torngat the permanent snowline lies at about 3,500 feet. Widespread lowering of the snowline across Labrador-Ungava, resulting from progressive climatic deterioration at the onset of glacial times, would rapidly place an immense area of the plateau above the snowline, and hence form a huge reservoir of névé long before any significant accumulation could occur in the Torngat. Although many more data are needed, it is concluded that at the onset of Wisconsin times glacierization across a large area of the Labrador-Ungava plateau would be instantaneous with any significant lowering of the snowline. Furthermore, although small glaciers would develop concurrently in the Torngat, the build-up of an ice cap on the plateau to the southwest would starve them of precipitation and the first major glaciers would move from the west and southwest through the mountains to the Atlantic Ocean.

Conclusions

1. The highest summits of the Torngat Mountains were completely submerged by eastward moving continental ice at the height of the Wisconsin Glaciation. The ice reached a thickness of possibly 1,000 feet above the «central» summits and the highest coastal summits were probably covered by at least a thin cover when the ice margin reached the vicinity of the outer edge of the continental shelf, some 90 to 100 miles east of the present coastline.

2. Local glaciers in the Torngat never reached significant dimensions, and in late-Wisconsin times most cirque glaciers did not even overspill the cirque thresholds.

3. The final movement of ice, excluding the local movement within the cirques, was from the west, and during the final stages of the Wisconsin Glaciation there occurred the rapid melting in situ of thick masses of ice.

4. Two, and possibly three separate glacial periods are recognized from the morphology of the area.

5. Instantaneous glacierization of a large area of the Labrador-Ungava plateau is considered the most likely method of initiation of a continental ice sheet in north-eastern North America at the onset of each glacial period.

Author's Note

Further field work during the summer of 1957 leads to the conclusion that there was a major discontinuity between the period of complete inundation of the high summits by ice and the period when the surface of the ice stood at the 2,000 to 2,200 foot level. This implies that two distinct glaciations, separated by an interval of complete or partial deglacierization occcurred in late Pleistocene time.

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AVALANCHE CLASSIFICATION

M. R. DE QUERVAIN
Swiss Snow and Avalanche Research Institute Weissfluhjoch/Davos

Abstract

The problem of avalanche classification is much older than that of snow classification. Nearly every scientist, alpinist or alpine writer dealing with avalanches has proposed a separate system.

The author of this note wishes to give a survey of various existing avalanche classifications, including a suggestion which has been published by R. Haefeli and himself. Whether or not an effort should be made for setting up a standard classification may be decided by the commission.

In the course of the last 100 years a great number of avalanche classifications have been suggested. There is a preference in certain countries or among certain groups for the one or the other system but hardly any one has fully been adopted. Although the matter is not of primary importance it is proposed to discuss it once on an international level. Would it be useful and possible to agree on a standard classification, either by accepting one of the existing systems or by creating a new one? An excange of opinions on this question seems to be desirable in any case.

No doubt, the project would be opposed by some serious difficulties:

— Not all languages dispose of a suitable vocabulary in this range.

— Certain terms which are not well defined and therefore objected by many authors are already deeply rooted in the common language.

— A classification may suit a scientific purpose but not necessarily a general application and vice versa.

For approaching the problem, the following procedure is recommended: First the general need and the particular interests of science, technique and mountaineering etc. in a classification should be established. Next all existing classifications have to be examined. It then may appear if any of them is satisfactory or if it is necessary and promising to try a new system. In this paper an attempt in this direction is made, consisting in an analyis of the requirements to be fulfilled of a classification and in a survey of existing classifications (without discussion).

1. Requirements of an avalanche classification

a) *General*

A classification, whether made for scientific use or for popular application, should be scientificaly correct and based only on well-established facts and not on theories. It should be able to survive any new results of research. New facts, however, may complete and refine the classification if necessary. A close agreement should be obtained in the terminology of the different languages.

b) *Scientific and technical classification*

The scientist wants to classify the entire range of facts without regard to practical application. There are two principal methods of classifying in science. A *deductive* scheme proceeds from an universal term to the special individual by subdivision

(for example the classification of animals with species and sub species). An *additive* scheme builds up the general term by combining the single independant characteristics. Mixed deductive and additive schemes are often used. Which system is best for an avalanche classification may be decided after the analysis is made. As an avalanche is not only a steady *object*, which can be photographed, but also an *event* with its antecedents, particular difficulties arise in building up a consistent classification.
Objects of an avalanche are:
— the terrain with its configuration, exposition and coverage
— the snow cover of a certain shape, depth and stratification
— type and dimension of the fracture of the snow
— momentaneous appearance of the movement (for example a snow cloud)
— the resulting change of the snow cover along the track of the avalanche including the deposit. Changes at other objects (destructions)
The *event* is characterized by:
— development of weather in advance (snow fall, wind, temperature)
— starting incident (spontaneously or foreign influence)
— development of movement and dimensions
— type of movement (sliding, rolling, turbulent)
— velocity
— dynamic effects

The events are facts as well as the objects, but in general there is not as much known about them.

The technical requirements of an avalanche classification are not very different from the scientific ones. Permanent avalanche protection is the main technical problem. For this avalanches have to be characterized according to fracture type, dimensions, movement and effects from the starting point to the end of the deposit. As not only scientifically educated people have to deal with technical aspects, the classification must be simple. In technical work (mapping, statistics) code systems or graphic symbols for representing the date are often useful and should also be investigated.

c) *Classification for mountaineering*

The mountaineer is interested in avalanches for reasons of safety mainly. He wants to know where and when an avalanche may occur, and, if it does, what risk is involved. With regard to rescue operations some knowledge on the general behaviour of avalanches is also desirable of course.

Thus, a classification for mountaineering purpose should above all deal with the conditions of avalanche formation i.e. with the terrain and the snow quality with respect to the foregoing weather. A simple and concise scheme and intelligible terms are also important requests.

2. Examples of existing avalanche classifications

These examples, originating from european countries only, are not given in all details, nor do they represent a complete survey. The English translations from the original languages made by the author may be susceptible to improvement. In brackets the original wording is added if necessary

a) *F. Ratzel*, Germany (1889) distinguishes between the classical terms

—dust avalanches (Staublawinen)

— ground avalanches (Grundlawinen)

«Ground avalanches» are defined as avalanches, dragging along soil and rocks form te ground. He proposes to denominate the form of movement by
— rolling avalanches (Rollawinen) and
— sliding avalanches (Rutschlawinen)

b) *J. Coaz*, Switzerland (1888), defines avalanches as «big snow slides» (grosse Schneeschlipfe). He also uses
— dust avalanches (Staublawinen)
— ground avalanches (Grundlawinen)

The second type is characterized as snow stream of low velocity intermingled with grass and rocks.

C. is not in favour of «rolling» and «sliding avalanche». Glacier avalanches are also mentioned as a third type.

c) *V. Pollak*. Austria (1891)

1. Dust avalanches ((Staublawinen).
— pure dust avalanches (air borne)
— common dust avalanches (partly air borne, partly loose ground avalanche)
2. Ground- or mass-avalanches (Grund- oder Massenlawinen)
3. Surface layers avalanches (Oberlawinen).

The ground avalanche is defined as an avalanche flowing along the ground. Small avalanches are also called «snow-slides» (Schneerutsche).

d) *A. Allix*, France (1925) gives a two dimensional system of the following kind (not complete)

form of movement	location	kind of snow: dry	kind of snow: wet
	at start (départ)	flowing (coulante)	—
		rolling (roulante)	rolling
		sliding (glissante)	sliding
	in the course	cascade	cascade
		— whirl (trombe)	—
		— jet (fontaine)	—jet
		flow (coulée)	flow
		— superficial	— rapid (rapide)
		— to ground (de fond)	— slow (lente)
	at end (arrivée)	cone	
		tongue (langue)	

The term ground avalanche (avalanche de fond) is transformed into « coulée de fond ».

e) *F. Fankhauser*, Switzerland (1929) considers the classical distinction dust avalanches — ground avalanches not to be satisfactory.

He reduces the system to
— dry avalanches
— wet avalanches
— mixed avalanches

f) *G. Seligman*, England (1936)
I. Dry snow avalanches
1. dry powder-snow avalanches
a) wild-snow avalanches
b) dry new-snow avalanches
c) dry settled-powder avalanches
2. dry old-snow avalanches
II. Wet-snow avalanches
1. damp-snow avalanches
2. medium wet snow avalanches
3. very wet snow avalanches
III. Wind slab avalanches
IV. Ice avalanches

The old terms of « Staublawine » and «Grundlawine » «have gone for good» according to Seligman.

g) *W. Paulcke*, Germany (1938) thinks the quality of the snow to be essential for the classification.

Dry snow avalanches	(Trockenschneelawinen)
a) wild-snow avalanches	(Wildschneelawinen)
b) dry loose snow avalanches	(trockene Lockerschneelawinen)
c) packed snow avalanches (snow shield avalanches)	(Packschnee-Lawinen) (Schneeschild-Lawinen)
d) compressed snow avalanches (snow slab avalanches)	(Presschneelawinen) (Schneebretter)
e) depth hoar avalanches (dry old snow avalanches)	(Schwimmschneelawinen) (trockene Altschnee-Lawinen)
B. Moist and wet snow avalanches	(Feucht- und Nass-Schneelawinen)
f) moist and wet loose snow and packed snow avalanches	(feuchte, nasse Lockerschnee- und Packschneelawinen)
g) wet firn snow avalanches (wet old snow avalanches)	(nasse Firnschnee-Lawinen) (nasse Altschneelawinen)
C. Glacier avalanches	
h) firn-ice avalanches	(Firneis-Lawinen)
i) glacier-ice avalanches	(Gletscherlawinen)

h) *W. Flaig*, Austria (1955)

After having published other suggestions, he gives in his recent book the following scheme:

I. Loose snow avalanches	(Lockerschneelawinen)
a) dry loose snow avalanches	(trockene Lockerschneelawinen)
— *powder avalanche (following the ground)	(*Pulverlawine)
— *dust avalanche (air borne)	(*Staublawine)

b) wet loose snow avalanches	(nasse Lockerschneelawinen)
— *snow slides, firn avalanches	(*Schneeschlipfe) (Firnlawinen)
— *ground avalanches	(*Grundlawinen))
II. Solid snow avalanches	(Festschneelawinen)
a) dry solid-snow avalanches	(trockene Festschneelawinen)
(*snow slab avalanches)	(*Schneebrettlawinen)
b) wet solid-snow avalanches	(nasse Festschneelawinen)
(*snow-cloth avalanches)	(*Schneetuchlawinen)
III. Mixed avalanches	
*stroke avalanche	(*Schlaglawine)
(a combination of various kinds of snow)	

*popular denomination

i) Draft of *Snow and Avalanche Research Institute* with *R. Haefeli*, Switzerland (1955)
Snow avalanches: dislocation of a snow mass over more than 50 m length of track
Snow slide : dislocation of less than 50 m length of track

criterion	alternative characteristics and denomination	
A type of rupture	starting from a line slab avalanche (Schneebrettlawine)	starting from a point loose snow avalanche (Lockerschneelawine)
B position of sliding surface	above the ground inside snow cover	on the ground
	surface layers avalanche (Oberlawine)	full depth avalanche (Bodenlawine)
C state of humidity	dry snow	wet snow
	dry snow avalanche (Trockenschneelaw.)	wet snow avalanche (Nassschneelawine)
D form of the track	open even track	channelled track
	unconfined avalanche (Flächenlawine)	channelled avalanche (Runsenlawine)
E form of movement	whirling through air	flowing along ground
	airborne powder avalanche (Staublawine)	flowing avalanche (Fliessslawine)

The english translations of the avalanche types have been proposed by the British Glaciological Society, the British Alpine Club and the Alpine Skiing Club. With this kind help no decision has been anticipated, of course, about adopting the system by the mentioned bodies.

The german terminology is based on some well introduced denominations. However, the term «ground avalanche» (Grundlawine) could not be incorporated in the scheme. It is to wide for any of the single characteristics, and is used as can be seen from previous classifications, mostly for describing a combined type, namely a heavy wet and dirty spring avalanche, following a well known track. In this sense the term may be carried on.

SNOW CLASSIFICATION

M. R. DE QUERVAIN
Weissfluhjoch Davos

It is reminded, that at the Oslo congress 1948 the need for a standard snow classification was urged. At the Bruxelles congress 1951 a draft classification was presented by a three men committee and adopted by the commission of snow and ice, provided that this classification would be approved by the World Meteorological Organisation. In the negociations with WMO certain discrepancies were established. It appeared that WMO also was revising the terminology of precipitation (part of the coud atlas). In 1956 the author of this note was designated by he Association of Hydrology to discuss the two classifications in question with a representative of WMO (Prof. Bleeker) in order to get them adjusted. The result of a conference held in Geneva has been communicated to the Secretary of the Association of Hydrology and the President of the Commission of snow and ice. It is suggested to bring it before the commission for final approval.

GEOCRYOLOGY AND ITS MAIN PROBLEMS IN THE USSR

P. F. SHVETSOV

1. For the last thirty years (1927—1956) a new branch of geologo-geographical and engineering-geophysical sciences-geocryology, has considerably developed in the Soviet Union. It studies:

a) laws of freezing and thawing of the Earth's crust, formation development and propagation of frozen soil zones and rocks (cryolite zones) under various physico-geographical and geological conditions;

b) characteristics of composition, structure and physico-mechanical properties of freezing, frozen and thawing soils and rocks;

c) geophysical, physico-geological, geomorphological and hydro-geological phenomena associated with processes of freezing, and thawing of soils and rocks;

d) a control of the effect of freezing, frozen and thawing soils and rocks upon construction, agriculture, etc.

2. A perennial cryolite zone (permafrost) cannot be treated as a result of the climate only. Physico-geographical and geological conditions are of serious importance for the formation and development of the permafrost.

3. The chief objective of geocryology is the study of processes of heat exchange between lithosphere, soil and atmosphere, as well as between ground and buildings, in order to work out the best types of firm constructions.

4. Besides these problems, the Soviet geocryologists also study the following:

a) the distribution of frozen soils and rocks;

b) mechanical interaction between a building and freezing, frozen and thawing ground and methods of controlling it;

c) physics of the cryogen processes.

О СОДЕРЖАНИИ И ОСНОВНЫХ ПРОБЛЕМАХ ГЕОКРИОЛОГИИ В СССР

П. Ф. Швецов.

За последние 30 лет (1927-1956 гг.) в Советском Союзе сформировалась и значительно развилась новая отрасль геолого-географических и инженерно-геофизических знаний — геокриология. Истоками ее являются результаты исследований зоны мерзлых горных пород на севере и востоке Евразии, произведенных А. Ф. Миддендорфом, А. И. Воейковым, Л. А. Ячевским, В. А. Обручевым, Б. Б. Полыновым, А. В. Львовым, А. А. Григорьевым и М. И. Сумгиным — создателем и руководителем школы советских мерзлотоведов. Особенно значительна роль крупнейшего русского геофизика А. И. Воейкова в формировании теоретических основ геокриологии; он выработал первые начала учения о теплообмене в системе литосфера-почва-атмосфера (1886, 1889, 1904).

В 1927 г. появилась первая в мире стройная и полная сводка данных о распространении, температурном режиме и мощности « вечной мерзлоты » (многолетней криолитозоны) и представления о происхождении и закономерностях развития ее (М. И. Сумгин); она была развита на основании результатов исследований и изысканий площадок и трасс под новостройки первой и второй пятилеток и дополнена в фундаментальном труде советских мерзлотоведов « Общее мерзлотоведение » (М. И. Сумгин, С. П. Качурин, Н. И. Толстихин и В. Ф. Тумель, 1940). В 1937 г. были созданы « Основания механики мерзлых грунтов » (Н. А. Цытович, М. И. Сумгин), а в 1941 г. основы учения о подземных водах « мерзлой злны литосферы » (Н. И. Толстихин).

К этому же времени были разработаны и проверены производственным опытом принципы и способы :

1) предохранения низкотемпературных мерзлых грунтов (среднегодовая температура не выше —0°,5 С) от протаивания и осадок под тепловыделяющими сооружениями путем устройства проветриваемых зимой подполий;

2) строительства водопроводов в мерзлых грунтах (среднегодовая температура не ниже —1° С), надежная эксплуатация которых обеспечивается не только подогревом воды, но и особым гидравлическим режимом.

В период 1942-1956 гг. созданы научные основы :

1) механики мерзлых и оттаивающих грунтов;
2) строительства сооружений на протаивающих в периоды возведения и эксплуатации их естественных основаниях;
3) расчета глубины промерзания почвагрунтов;
4) температурного районирования многолетней криолитозоны;
5) электрометрического (методом постоянного тока) определения элементов залегания и состава толщ мерзлых горных пород;
6) поисков и разведок подземных вод в районах многолетней криолитозоны.

За это же время сформировались современные представления о положении и динамике южной границы многолетней криолитозоны, основы структурного ледоведения — петрология континентальных льдов, теория происхождения наиболее широко распространенных залежей грунтового льда, первые представления о развитии и значении геохимических процессов в мерзлых горных породах.

Многое сделано уже в разработке и производственной проверке способов искусственного промораживания и протаивания грунтов.

Геокрилогия в ее современном виде представляется нам учением :

1) о закономерностях промерзания и протаивания земной коры, формирования, развития и распространения зон мерзлых почв и горных пород (криолитозон) в ходе теплообмена в системе литосфера — почва — атмосфера при определенных физико-географических и геологических условиях;
2) об осоеностях состава, строенния, сложения и физико-механических свойств промерзающих, мерзлых и протаивающих почв и горных пород;
3) о геофизических, физико-геологических, геомерфологических и гидрогеологических явлениях, связанных с процессами промерзания, протаивания и изменениями температуры почв и горных пород и
4) о приемах управления процессами теплового, влажностного и механического взаимодействия промерзающих, мерзлых и протаивающих почв и горных пород с результатами производственной деятельности людей (сооружениями, горными выработками, пашнями, культурными растениями, искусственными водоемами и др.).

В таком понимании содержания геокриологии (мерзлотоведения) нет ничего оригинального, принадлежащего одному лицу; оно составлялось значительной частью коллектива Института мерзлотоведения им. В. А. Обручева на протяжении 15-16 последних лет. В мерзлотоведении происходит сейчас примерно то же, что происходило в почвоведении в последней четверти XIX столетия : многие из нас все более настойчиво заявляют о необходимости и на словах выйти за пределы изучения состояния, свойств, мощности и распространения определенной геозоны, заняться иссл едованием состава, строения, происхождения и непрерывной смены состояний — развития ее. На деле мы уже шагнули далеко в этом направлении, под влиянием потребностей технического прогресса в строительном производстве в горнодобывающей промышленности.

Ко времени юбилейной сессии Института мерзлотоведения имени В. А. Обручева (декабрь 1945 г.), отмечавшей пятнадцатилетие существования академического научно-исследовательского учреждения по изучению многолетней криолитозоны, был уже поставлен (М. И. Сум гиным, И. Я. Барановым, В. А. Кудрявцевым, Н. И. Толстихиным и В. К. Яновским) вопрос о развитии, « динамике вечной мерзлоты ». Сама постановка этого вопроса требовала серьезного пересмотра важнейших положений мерзлотоведения о том, что « вечная мерзлота » — продукт климата и наследие ледниковой эпохи.

Действительно, при таких взглядах на происхождение и условия развития « вечной мерзлоты » она оказывалась чем-то внешним и посторонним по отношению к земной коре, навязанной литосфере сверху в период оледенения, результатом катастрофы. « Вечная мерзлота », проникшая в земную кору во время катастрофы — оледенения суши, могла будто бы развиваться только в одну сторону — деградировать вследствие потепления климата. Этим самым как бы не признавалось термодинамическое саморазвитие почвы и литосферы.

Против таких суждений, преуменьшавших возможности искусственного изманения температуры подпочвы в нужном для производства направлении,

вслед за М. И. Сумгиным и В. К. Яновским обоснован но и более решительно выступили В. А. Кудрявцев и И. Я. Баранов (1947-1951 гг.).

Многолетнюю криолитозону (вечную мерзлоту, Permafrost) нельзя рассматривать как продукт климата, понимая под последним то, что рекомендуют понимать многие известные климатологи (А. И. Воейков, Б. П. Алисов и др.). В 1951-1955 гг. советские геокриологи сформулировали положение : мерзлые зоны земной коры сформировались в результате и развиваются в ходе теплообмена в системе литосфера — почва — атмосфера (и мировое пространство) при определенных физико-географических и геслогических условиях. Климат — лишь одно из условий и, вместе с этим, продукт тепло- и влагообмена в указанной системе и в системе слои воды — атмосфера, при наличии адвекции.

В связи с новыми выводами о происхождении, условиях существования и развития мерзлой зоны литосферы, возникла настоятельная необходимость в уточнении истории знаний и понятий мерзлотоведов, в поисках соответствующих слов для более точного выражения понятий.

К достижениям геокриологии может быть следовало бы причислить и появление первых основ учения о теплообмене в системе литосфера — почва — атмосфера, которые не появились до сих пор ни из недр метеорологии, ни из недр геотермики. Первая из этих отраслей геофизики — метеорология исследует теплообмен подстилающей поверхности с атмосферой (и мировым пространством), в торая, т.е. геотермика — развитие температурного поля литосферы ниже слоя с переменными в году температурами. Тепло — и влагообмен слоя с сезонными изменениями температуры представляют собою весьма сложный, но зато и исключительно интересный с научной и практической стороны объект геофизики.

Учение о теплообмене в системе литосфера — почва — атмосфера основывается на принципах, теоретических и экспериментальных достижениях как термодинамики, так и учения о теплопередаче. Последнее, пополнившись принципом термовлагопроводности, развивается сейчас (как и прежде развивалось) в связи с решением инженерно-технических задач и в учреждениях технических наук.

Геокриологи физического и инжернерного направлений в ближайшие годы смогут, видимо, создать прочные основы учения о теплообмене в системе литосфера — почва — атмосфера, чем обогатят самое геокриологию, а также общую геофизику.

Главной и наиболее общей задачей геокриологии на современном этапе развития ее является всемерное углубление и расширение теоретических и экспериментальных исследований процессов и условий *тепло-и массообмена в системах литосфера — почта — атмосфера и грунт — сооружение*, для выявления новых способови оценки каждого из известных приемов обеспечения устойчивости сооружений и искусственного протаивания горных пород за счет естественных источников тепла в разных физико-географических и геологических условиях.

Кроме названной важнейшей проблемы советские геокриологи решают еще следующие три :

1) Основные закономерности и региональные особенности в развиитии и распространении мерзлых почв и горных пород. Сущность решения проблемы состоит в раскрытии закономерностей формирования и развития зон мерзлых почв и горных пород и их практического значения в пределах важных в народнохозяйственном отношении районов.
2) физика криогенных процессов и природа прочности мерзлых и оттаивающих грунтов. Знание физической сущности процессов промерзания и протаивания грунтов и закономерностей изменения показателей прочности их необходимо для решения задач тепловой мелиорации и основных теоретических задач обще и ниженерной геокриологии.
3) Изучение деформативно-напряженного взаимодействия сооружения с промерзающим, мерзлым и оттаивающим грунтом и управление этим взаимодействием. Намечается выявить основные закономерности работы льда и мерзлого грунта как среды, основания и материала различных сооружений.

Значительные шаги в решении этих проблем уже сделаны. После дние достижения советских геокриологов отражены в коллективном труде « Основы геокриологии », который будет, мы надеемся, опубликован в 1958 г.

GLACIOLOGICAL STUDIES ON AUSTERDALSBREEN, NORWAY, 1955-57

Members of the Cambridge Austerdalsbre Expedition

Abstract

A glaciological programme has been started on Austerdalsbreen, one of the distributary glaciers from Jostedalsbreen, Norway, with the aim of investigating ice flow at the foot of a large ice fall and the process of ogive formation. In 1955 gravity measurements of glacier depth were made; we also excavated a tunnel in the ice at the foot of the ice fall and surveyed its subsequent deformations. In 1956 detailed measurements were made of the deformation of the glacier surface, and a pipe was inserted into the ice in order to measure glacier flow at depth. We are continuing these studies in 1957. This paper reports the present state of the investigation and gives some preliminary results obtained from the surface deformation programme.

Introduction

When a glacier plunges down an ice fall, it produces at the bottom a series of undulations or waves running across the glacier. Further down stream these undulations give place to light and dark bands, which, as a result of the differential flow of the glacier, get pulled out into the well-known ogives or Forbes bands, noticed over a century ago by J. D. Forbes and used by him to demonstrate the existence of differential flow. Recent developments in the theory of glacier flow have helped us to understand many of the features exhibited by glaciers, but the undulations and the light and dark bands are not yet satisfactorily explained. One of the principal aims of the Cambridge Austerdalsbre Expedition has been to find out as much as possible about the nature of these features and their mode of formation.

For this purpose we need to study the shape and size of the undulations and their geometrical relationship to the light and dark bands. We also need to measure the surface strains and the ablation of the ice in the regions where the undulations are formed, where they mature and where they die away. If the surface strains are to be interpreted to the best advantage, it is also most desirable to know how the glacier is deforming at depth.

Work of the Expedition, 1955

The glacier selected for study was Austerdalsbreen, one of the distributary glaciers from Jostedalsbreen, the large ice cap in Western Norway. Two large ice falls known as Odinsbreen and Thorsbreen drop some 700 m. from the ice cap to a low valley. The glacier which they then form (Austerdalsbreen) shows all the phenomena of undulations and Forbes bands very well, and is also reasonably easy of access.

The main results of the 1955 field season have already been published ([1]) ([2]). First, the depth of ice in the more regular part of the tongue was determined by a gravity survey. This showed that the thickest part of the glacier, of depth about 100 m., lay near the snout; nearer the ice falls the depth fell to about 80 m. Second, a tunnel, 44 m. long, was dug horizontally into the foot of Odinsbre ice fall (where the surface slope was about 26°) to study the glacier flow at depth. We found that, as well as being carried forward by the glacier flow and being rotated, the axis of

the tunnel was being bent upwards, as in an active anticline; in addition the ice of the tunnel walls was compressing axially. The observation of bending shows that the strain-rates, and hence presumably the stresses, are not constant over the region of the tunnel; while the axial compression of the tunnel, together with strain measurements on the surface of the ice fall above the tunnel, show that a compressive stress must be acting along the line of flow of the glacier; its magnitude can be estimated as 3.1 bars.

Work of the Expedition, 1956

The work of the 1956 field season has not yet been published and will be briefly described here. The main projects were (*a*) the insertion of a pipe to measure glacier flow down to the bed of the glacier, and (*b*) a detailed study of the undulations and their motion.

(a) *Insertion of the pipe.* The pipe was inserted perpendicular to the surface, at the same region of the glacier as was used for the tunnel in 1955, that is, at the foot of the Odinsbre ice fall and at about the top of the region in which the undulations can be clearly identified. It was of aluminium alloy tubing of internal diameter 7.6 cm. (3 in.) and fitted with an electrically heated tip. When it had penetrated to a depth of 39 m. further heating had no effect and the tip is presumed to have reached rock. Whether this rock is the bed of the glacier is not certain, but rough calculations of the depth, based on the gravity depths measured lower down on the glacier tongue, make it reasonable to suppose that bedrock has been reached. The slope of the pipe at various depths was measured at the end of the 1956 summer season with a specially constructed inclinometer, and when a second set of such readings has been obtained it will be possible to compute the velocity profile of the glacier. Unfortunately a party which returned to the site in January 1957 was unable to find the pipe beneath the deep snow.

(b) *Study of the undulations and their motion*

(i) *Geometrical features.* The general appearance from a vantage point on the valley side is of two series of undulations, one from each of the two ice falls, which coalesce after about the first wave to form a single series. We made a special study of the series which starts at the foot of the Odinsbre ice fall. 7 troughs and 7 crests can be distinguished before the waves die away. The maximum amplitude, measured perpendicularly from the average surface of the glacier, is 5.8 m.; the wavelength from crest to crest diminishes from 180 m. to 75 m. as one goes down glacier; the initial wavelength is thus 32 times the initial amplitude. We may also note that, if the depth of the glacier is estimated by assuming that the pipe has reached bedrock, the initial wavelength is 4.6 times the depth of the glacier.

(ii) *Relation between wavelength and annual movement.* The summer velocity in the region of the undulations was measured by theodolite observations at a number of points both in 1955 and 1956. The annual movement of the glacier, calculated with the summer velocity, varies considerably from point to point but is nevertheless everywhere equal ($\pm$ 5%) to the wavelength of the undulations.

(iii) *Surface deformations.* Our main object was to study the role played by plastic deformation in the formation of the undulations, and in particular to test the theory which has often been put forward that the crests are caused by a longitudinal compression or buckling. For this purpose 23 stakes were placed in a row at intervals

of 30 m. down the main longitudinal axis of the wave system — starting in the lower part of the ice fall and extending through the first two and a half waves, a total distance of 630 m. The distances between successive stakes were measured by steel tape at four different times over a period of 28 days, and in this way a curve was obtained showing the longitudinal strain rate $\dot{\varepsilon}_x$ as a function of x, the distance down the glacier. In work of this type spurious strain-rates can arise from the reboring of the stakes (which was frequently necessary) and from the different inclinations to the vertical of the different stakes. We took great care to measure these effects and to allow for them in the calculations. By internal checks the accuracy in $\dot{\varepsilon}_x$ is estimated at $\pm$ 0.01 yr^{-1}.

The broad result is that in the area covered by the stakes there is an average longitudinal compressive strain-rate of the order of 0.5 yr^{-1}, but, more interestingly, the strain-rate is very far from uniform over the area. Its highest value, 1.57 yr^{-1} (compressive), occurs at the topmost stake interval. At a point only 120 m. lower down the glacier the strain-rate has fallen to 0.03 yr^{-1} (compressive). As one moves further down glacier the strain-rate oscillates several times, rising to 0.80 yr^{-1} (compressive) and falling once more to 0.02 yr^{-1} (tensile) before levelling off to 0.31 yr^{-1} (compressive) at the lowest point of the line of stakes. The reason for this behaviour is discussed in section (vi) below.

(iv) *Velocity distribution in the wave system.* We measured the vertical velocities, both absolute and relative, of all the stakes in the wave system, and, by combining the measurements with those already described, we have calculated the absolute horizontal and vertical velocity components of all the stakes. It appears from these measurements that the upper layers of the ice move along comparatively smooth streamlines, which, when seen in profile, are deflected only slightly, if at all, by the undulations of the surface. In other words, the stream lines at the surface are much smoother than the surface itself.

(v) *Strain-rate tensors.* In addition to the above measurements on our main line of stakes we made complete determinations of the rate of strain tensor at five points on the main line. By placing stakes at the four corners and at the centre of each of five squares and measuring the subsequent distortions, the principal strain-rates were found to within $\pm$ 0.004 yr^{-1} and their directions to within $\pm$ 0.4°. In all five cases the maximum tensile strain-rate was perpendicular to the average direction of the crevasses present, in agreement with theory.

(vi) *Theory of the strain-rate distribution.* If the crests are formed by a longitudinal compression or buckling, we ought to find that the longitudinal compression rate is high at the crests and low at the troughs. No such correlation exists in our measurements. Nor are the crests rising and the troughs falling relative to their surroundings. We can only conclude, then, that a buckling process, if it occurs at all, does not take place in summer at the foot of the Odinsbre ice fall.

Nevertheless, as mentioned above, we do observe very considerable oscillations in the longitudinal strain-rate over the wave area, and these oscillations have been a most puzzling feature of the investigation. The following theory, which will be described more fully elsewhere, now seems to offer a consistent explanation of the observations.

When a glacier moves over a part of its bed whose curvature is changing the ice must bend about a horizontal axis. If the bed becomes more convex, the upper layers of the glacier will extend longitudinally and the lower layers will compress. If the bed becomes more concave the opposite strains will occur. The important points are, first, that this effect depends on the rate of change of the curvature $\varkappa$

of the bed, that is, on $d\varkappa/dx$, and not on $\varkappa$ itself, and, second, that the strain-rate $\dot{\varepsilon}_x$ from this cause changes sign through the thickness of the glacier. A theory has already been developed ([3]) which gives the dependence of the longitudinal strain-rate $\dot{\varepsilon}_x$ on $\varkappa$ itself and on the rate of ablation (equation 36). The strain-rate from this cause, unlike the strain-rate arising from $d\varkappa/dx$, is uniform through the thickness of the glacier. In the derivation in ref. ([3]) it is postulated that $d\varkappa/dx$ is small. It is now proposed to add to the theoretical expression for $\dot{\varepsilon}_x$ a new term, proportional to $d\varkappa/dx$ and based on the above considerations, which will take account of cases where $d\varkappa/dx$ is not small. The proposed expression, which will not be fully derived there, is as follows:

$$\dot{\varepsilon}_x = u\varkappa \cot\alpha + \frac{1}{2} hu \frac{d\varkappa}{dx} + \frac{\Phi'}{h} - \dot{\varepsilon}_z ,$$

where ε_x is the longitudinal strain-rate at the upper surface of the glacier (positive when tensile), u is the mean velocity of the glacier, $\varkappa$ the curvature of the bed (positive when convex), α the slope of the surface, h the depth, Φ' the rate of accumulation (negative if ablation) and $\dot{\varepsilon}_z$ the transverse strain-rate (positive when tensile).

To test this expression on the present observations it is necessary to know the rate of change of curvature of the bed. On the present theory the form of the bed is closely similar to that of the streamline at the upper surface, and fortunately we have measured this. All parameters in the expression for ε_x are therefore known, and one may compare directly the theoretical distribution of $\dot{\varepsilon}_x$ with that observed. Considering that there are no adjustable parameters in the theory the result is very satisfactory — the general form of the two curves is similar, and the maxima and minima of the observed curve are all present in the correct positions (within 15 m. except for one case) in the theoretical curve. The magnitudes of the oscillations in the theoretical curve agree with those observed to within a factor of 2. One sees from the comparison that the observed longitudinal strain-rate is predominantly due to two terms: a steady compressive term of magnitude about 0.4 yr^{-1} arising from the general concave nature of the bed at the foot of the ice fall, and an oscillatory term of magnitude about $\pm$ 0.2 yr^{-1} which arises from $d\varkappa/dx$ as just discussed.

The essential content of this part of the work may be epitomised by saying that we have established a connection, both experimental and theoretical, between the form of the surface streamline and the distribution of $\dot{\varepsilon}_x$.

The position we have now reached in the argument is that the strain-rates at the foot of the icefall have been measured and found to be irrelevant to the process of wave formation; they are in fact produced predominantly by the form of the bed. We must therefore search for the process of generation of the waves elsewhere in position or time, and, in particular, we are led to enquire about conditions higher up the icefall.

(vii) *Velocity distribution in the ice fall.* During the 1956 field season photographs of the Odinsbre ice fall were taken twice daily. The series of still pictures so produced has subsequently been made into a cine-film to show the movement of the ice, and, by direct measurement of the photographs the velocity distribution in the ice fall in summer has been roughly determined. We also have the theodolite measurements of velocity at the foot of the ice fall mentioned in (ii) above. It appears that the summer velocity of the ice increases from 640 $\pm$ 100 m/yr at the top of the ice fall to 1100 $\pm$ 100 m/yr in the middle, and then decreases to 400 $\pm$ 15 m/yr near the bottom. The ice is thus extending longitudinally in the upper part of the fall and compressing longitudinally in the lower part. This picture of the motion accords with the appearance and location of the seracs in the ice fall.

The very high velocity measured in the ice fall means that a piece of ice starting

from the top would be nearly at the bottom after one year, and after 1.5 years it would be at the position occupied in August 1956 by the first trough at the foot of the ice fall. The discovery of this high velocity has led one of us to propose the following theory for the generation of the waves.

(viii) *Theory of wave generation.* The net annual ablation of a glacier at the firn line is zero. Below the firn line the net annual ablation increases and, according to the usual picture of glacier motion, and if the glacier is in a steady state, the velocity of the ice normal to the glacier surface is just such as to keep the surface in a fixed position. In this picture one is usually able to assume, with negligible error, that the net annual ablation only changes slightly in a distance equal to an annual forward movement of the ice. In the present case, however, this last condition is far from being satisfied. The firn line is situated at the top of the ice fall. According to our measurements of summer velocity, the ice at the firn line moves in one year a distance of 900 m. down the glacier surface and at the same time drops 460 m. in altitude. The net annual ablation over this distance changes from zero (at the firn line) to about 5 m. This being so, the sections of the glacier which pass the firn line at the beginning of winter and the beginning of summer are exposed to very different melting (and other) conditions. It is not necessary to make any more detailed analysis to conclude that surface undulations should be formed. Their spacing should be everywhere equal to the annual movement and their initial amplitude should be comparable with the net annual ablation. Both conclusions accord with our observations.

Continuing this line of thought the following picture of the processes at work has been put forward, with the reservation that, since the experimental and analytical work is still in progress, conclusions must, for the time being, remain provisional. In summary, then: (a) The undulations are formed by the abrupt change of ablation which the glacier suffers as it moves down the ice fall, as described above. (b) The undulations are then accentuated by the general longitudinal compression in the lower part and at the foot of the ice fall. (c) At the same time the waves tend to flatten themselves by preferential melting of the crests. This effect shows up very clearly in our measurements: the summer rate of melting of the first few crests is 30% higher than the rate for the troughs.

(ix) *Relation of the undulations to the light and dark bands.* The light and dark Forbes bands become distinct and measurable precisely at the region where the surface undulations become indistinguishable. We have tried to find the geometrical relation between them by taking photographs of the Forbes bands from high up the valley walls, with control marks on the ice itself. The dark bands appear, by extrapolation, to emanate from points which are slightly (0.1 ± 0.2 of a wavelength) down glacier from the crests.

Other Activities

In addition to the work outlined above, several other related programmes have been carried out, either by the Cambridge expedition or by associated expeditions from Nottingham University, Brathay Exploration Group and elsewhere. Measurements have been made at points where ice runs close to the rock wall without intervening moraines, to determine the rate of flow of the ice past the rock. The velocity is quite large, being 100 m/yr only 1.2 m. from solid rock at one point. This is an appreciable fraction of the velocity of the ice in the centre of the glacier.

Measurements of snow banding in crevasses have been made in the accumulation area of the glacier, and maps have been made of the glacier and of the moraines below its snout. A levelling survey has been made up the valley from Veitastrondsvatn, a lake into which the outflow water of the glacier runs some 14 km. below the snout, and echo-soundings have been made in this lake to discover its bottom profile. These other projects will be reported in detail elsewhere; one paper has already been published on deformation of ice beneath boulders ([4]).

The activities of the Expedition are not yet finished. It is hoped to make further measurements on the stake system and on the pipe in the summer of 1957. For this reason we emphasize once more that the conclusions drawn from the behaviour of the stake system must be regarded as tentative, and subject to modification in the light of further data.

Acknowledgments

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RADIATION MEASUREMENTS ON THE GREENLAND ICE CAP

M. DIAMOND and R. W. GERDEL
U. S. Army Snow, Ice and Permafrost Research Establishment

Abstract

Measurements were made of global and net radiation during the summer of 1955 at a site on the Greenland Ice Cap located near 78° N. latitude and at an elevation of 6800 ft. Only 15% of the total incident global radiation was absorbed by the snow cover and diffuse sky radiation amounted to only 19% of all incoming global radiation. Incident global radiation was reduced by 6% in the presence of a 0.5 cloud cover and under full overcast conditions the snow surface received 65% of the global radiation measured on clear days. The long-wave radiation balance was always negative.

The heat balance of the snow cover during the 1955 summer as computed from the radiation measurements and a temperature profile in the snow was found to be 7.6 ly/day at the Ice Cap site.

Introduction

North of 75° the sun is continuously above the horizon from the first of May to the last of August. In June and July the Arctic has a potentially greater supply of solar radiation than any other area in the world. The reflection of radiation from the snow and ice cover, which may amount to more than 85% of the incident radiation, decreases the solar energy available for melting, thus aiding in the perpetuation of the extensive ice caps and snow fields. On the other hand the large amount of reflected radiation may provide an additional supply of thermal energy for applied use in the Arctic.

Any efforts to apply a heat budget balance to the problem of distribution, accretion, and ablation of the snow cover, ice caps, or glaciers requires quantitative information on solar and terrestrial radiation. Reliable information on available solar radiation is also needed as a basis for determining the applicability to the Arctic of some of the promising recent developments in applied use of solar energy.

The high ice cap in north-central Greenland provides an almost ideal Arctic site for radiation measurements during the summer months. The level snow surface extends circumferentially to the horizon. The almost constant wind movement of snow maintains a surface with a high albedo. The persistent low temperature and absence of melting at this location eliminates some of the problems associated with the measurement of radiation and the derivation of a radiation heat budget in a more temperate climate. These advantageous conditions were utilized during the summer of 1955 to obtain a series of radiation measurements which provided information on the following parameters:

1. Global radiation (combined solar and diffused sky radiation) on a horizontal surface.
2. Global radiation hemispherically reflected from the snow surface (albedo).
3. Diffuse sky radiation with the solar disk occulted.
4. Incident solar radiation on a surface normal to the sun.
5. Net all-wave radiation in the 0.3μ to 300μ range.

Instruments and recorders were installed during the summer of 1955 at a permanent military station located on the Greenland Ice Cap approximately 200 miles east of Thule, near 78° N. latitude. The elevation of this station is about 6800 ft. A light-weight tubular aluminum tower about 8 m high was erected to support meteorological equipment, including some of the radiation-measuring instruments used in this study. Other radiometers and pyrheliometers were mounted on 4-ft high supports in the vicinity of the tower.

Incident global radiation was measured with a 10-junction Eppley hemispherical pyrheliometer mounted with the thermopile target facing the zenith.

Reflected global radiation was measured with a 10-junction Eppley hemispherical pyrheliometer mounted with the thermopile target facing, and in plane parallel with, the snow surface.

Normal incident radiation was measured with an Eppley normal incident pyrheliometer installed on a manually operated altitude-azimuth mount which permitted adjustment of the thermopile target normal to the solar disk each time a measurement was made.

Diffuse sky radiation was obtained at appropriate intervals by momentarily occulting the sun so the target of the pyrheliometer used for global radiation received no direct solar radiation.

Net all-wave radiation, which is the difference between the total incoming global and long-wave radiation and the total reflected global and outgoing long-wave radiation was measured with an aspirated Gier-Dunkle net hemispherical radiometer.

Fig. 1 — Meteorological instrumentation at Site 2 where the radiation measurements were made.

Fig. 2 — One of the Eppley total hemispherical pyrheliometers (foreground) and the Eppley normal incident pyrheliometer (background), shown as installed at the Ice Cap site.

Fig. 3 — Gier-Dunkle radiometers for hemispherical measurement of total incoming all-wave radiation and net all-wave radiation, as installed at the Ice Cap site.

Recording of data

The hemispherical pyrheliometers used for incident global and reflected radiation measurements and the net radiometer used for all-wave radiation measurements were connected to recording potentiometers to provide a continuous strip chart record. Measurements with the normal incident pyrheliometer were made several times each day using a manually operated potentiometer.

Photographs of the installation at the Ice Cap research site are shown in Figures 1, 2, and 3.

Results

Incident and reflected global radiation and net all-wave radiation as tabulated from the recorder charts, are included as an appendix to this report. These data served as the basis for the development of those sections of this report which deals with specific phases of the radiation investigations.

Global radiation

The total daily incoming global radiation and the amount absorbed by the snow surface, computed from the difference between the incident and reflected global radiation are plotted in Figure 4. For comparison purposes the mean daily cloud cover is included in this figure. The total global radiation incident on the snow surface from 6 July to 7 August 1955 inclusive was 20, 628 ly*, and of this amount only 3059 ly, or about 15%, was absorbed at the snow surface. At the end of the dark winter period, the temperature of the snow cover was below —30°C and the atmospheric temperatures were consistently below freezing. Therefore, the small gain of heat from global radiation during the entire daylight period, even if there were no loss by long-wave back radiation, would be insufficient to permit any ablation at

* 1 langley (ly) = 1 cal/cm^2.

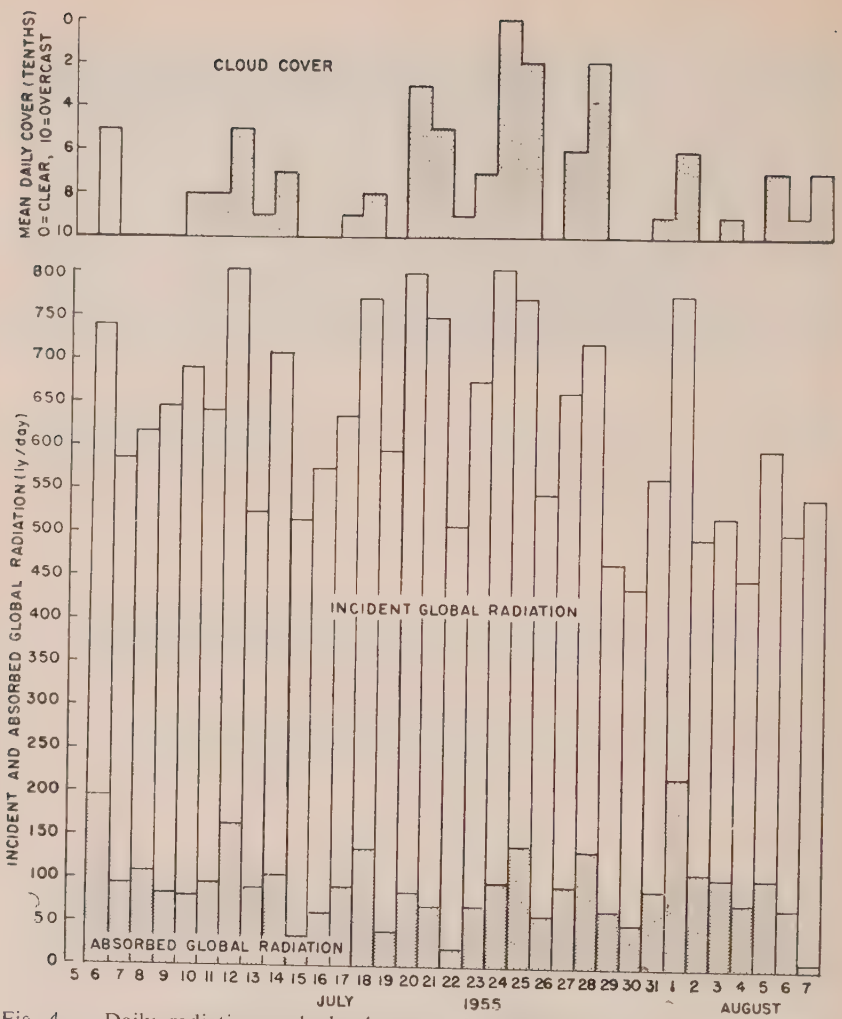

Fig. 4 — Daily radiation and cloud cover, Greenland Ice Cap, at 6800 ft msl. July-August, 1955.
Total incident global radiation on snow cover, 6 July - 7 August inclusive, was 20,628 langleys. Total global radiation absorbed by snow cover, 6 July - 7 August inclusive, was 3059 langleys.

this site, except for a minor amount of direct sublimation to the air. As shown later (see «Heat balance of the snow cover»), not more than 8% of the 3059 ly net gain in global radiation, or about 1% of the total incident global radiation, was retained by the surface layers of snow during the 33-day period covered in this study.

Albedo of Snow

The amount of incident solar radiation which is absorbed by a snow cover depends on the reflectivity or albedo of the snow surface. Some of the factors which

affect albedo are surface roughness, melting, refreezing of melt water, and the presence of contaminants. Most of these factors are associated with the age of the snow surface since the time of the last storm. The snow cover in the central portions of the Greenland Ice Cap is not subject to melting or contamination, and because of continuous drifting the surface is always relatively new. The continuous drifting and erosion produces a patterned surface which does appear to affect the albedo.

The albedo, A, of a surface is defined as the ratio between the reflected radiation, I_R, and the incident radiation, I;

$$A = I_R/I \tag{1}$$

and the radiation absorbed by the snow cover, I_a, is

$$I_a = I\,(1 - A). \tag{2}$$

Since albedo is not measured directly, errors in the observed values of I and I_R will yield incorrect values of R. Hubley (1955) observed a diurnal varation in the albedo of a melting snow cover, which he ascribed in part to changes in sun angle. Eckel and Thams (1939) found no dependence of the albedo of the winter snow on solar altitude. In a later study, Prohaska and Thams (1940) found that the diurnal variation of albedo was largest with air temperatures above freezing and almost negligible with air temperatures below freezing.

During the period 6 July to 7 August 1955, when albedo measurements were made at the Ice Cap site, air temperatures were always below 0°C and there was no visible evidence of melting at the snow surface. Skies were mostly cloudy; there where only 4 days with the cloud cover less than 0.5. Drifting or blowing snow occurred almost every day.

The daily values for albedo computed for all the hours of record are presented in Figure 5 and the mean hourly values for albedo under cloudy and clear skies,

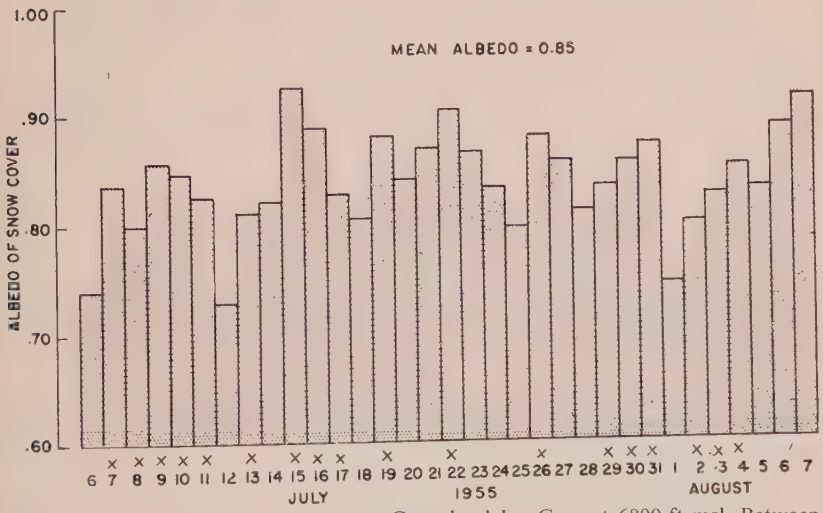

Fig. 5 — Daily albedo of snow cover, Greenland Ice Cap, at 6800 ft msl. Between 1000 and 1600 hr. 6 July - 7 August, 1955. X = Day on which snowfall occurred.

plotted against solar time are presented in Figure 6. On both clear and cloudy days, albedo was higher in the afternoon than in the morning (Table 1).

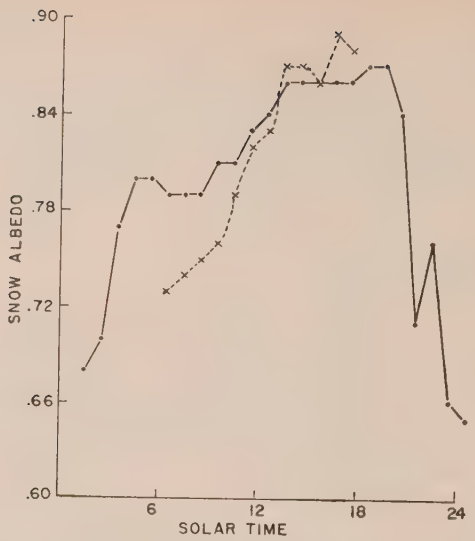

Fig. 6 — Variation of albedo with time, Greenland Ice Cap, 6800 ft msl.
. = Mean value for 30 days with more than 0.7 cloud cover.
× = Mean value for 3 days with less than 0.2 cloud cover.

TABLE 1

Mean Values of Albedo in the Morning and Afternoon

	Solar time 0500-1100	Solar time 1300-1900
Clear Day	0.77	0.87
Cloudy Day	0.80	0.86
All Days of Record	0.80	0.85

Note: On clear days, cloud cover was 0.2 or less, cloudy days had 0.7 or more cloud cover.

The difference may be due to reflections from the etched patterns on the snow surface, which were present more than 80% of the time during the period when the radiation measurements were made. The patterns are formed by wind erosion of the snow surface and frequently exhibit vertical or even undercut surfaces facing into the wind (Figures 7, 8).

At the Ice Cap site, the vertical surfaces faced toward the southeast, or into the morning sun.

The presence and orientation of erosional patterns on the snow surface may have sufficient effect on the albedo to require that specific measurements of reflected global radiation be made at each site where the precise determination of the energy

Fig. 7 — Variations in snow-surface patterns which may have a diurnal effect on albedo through change in sun angle and azimuth or by movement of the pattern as etching and filling takes place. Photographed near the research site on the Greenland Ice Cap.

balance is required. It appears also that the albedo of a permanent snow cover is much higher than that usually attributed to glacier snow and firn in the temperate zones.

Fluctuations in albedo which were recorded after 2100 and prior to 0500 hours, as shown in Figure 6, may be attributable in part to the instrumental problem involved in measuring the incident and reflected solar radiation at the low sun angles (of 3° — 18°) during these periods.

Diffuse sky radiation

Instantaneous measurements of diffuse sky radiation were made at frequent intervals by occulting the hemispherical pyrheliometer from direct solar radiation. The results are plotted in Figure 9. The diffuse sky radiation was also computed as follows: The solar radiation on a horizontal surface is computed from:

$$I_h = I_i \sin \phi \qquad (3)$$

where:

I_h = Solar radiation on a horizontal surface
I = Normal incidence solar radiation
ϕ^i = Sun angle (above horizon).

Fig. 8 — Sharply etched anvil-type drifts which may have a diurnal influence on albedo. Photographed near the research site on the Greenland Ice Cap.

The diffuse sky radiation is then computed from:

$$I_s = I_t - I_h \tag{4}$$

where:

I_s = Diffuse sky radiation
I_t = Total global (solar and sky) radiation
I_h = Solar radiation on horizontal surface

Both measured and computed values are listed in Table II. Values are similar except for the period of low sun angle (less than 15°) after 1800 hr, when the measured values were larger than the computed values.

At the Ice Cap site, diffuse sky radiation constituted 19% of the total incoming global radiation. In the temperate latitude, sky radiation has been estimated to be 30 to 40% of the incoming global radiation (King, 1912).* The low percentage of diffuse sky radiation at the high-elevation Arctic site may be attributed to low atmospheric turbidity since, as Linke (1942)** has shown, the ratio of sky radiation to global radiation is small when the atmospheric turbidity is low.

* Cited in Sutton (1953).
** Cited in Fritz (1951, p. 24).

TABLE II

Computed and Measured Values for Diffuse Sky Radiation

Date 1955	Time (hr)	Measured Sky Radiation (ly/min)	Computed Sky Radiation (ly/min)
27 July	1400	0.15	0.15
	1500	0.13	0.14
	1600	0.12	0.15
	1700	0.10	0.14
	1800	0.08	0.09
	1900	0.07	0.05
	2000	0.05	0.01
	2100	0.055	0.01
	2200	0.04	0
28 July	1200	0.14	0.14
	1400	0.16	0.13
1 August	1015	0.17	0.18

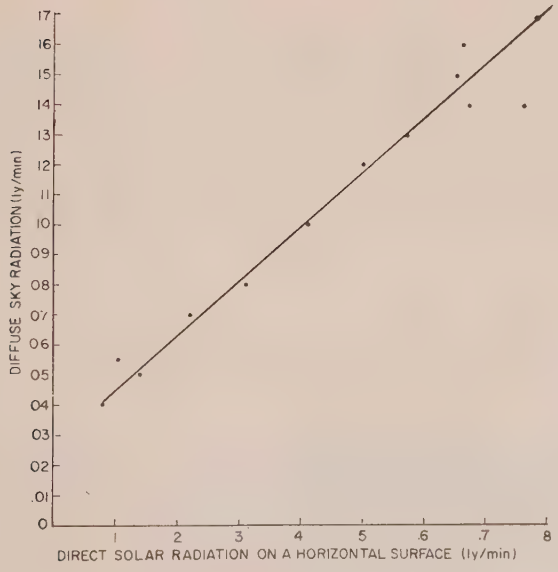

Fig. 9 — Diffuse sky radiation vs. direct solar radiation, Greenland Ice Cap, 6800 ft msl.

Effect of clouds on radiation

Observations of cloud cover were made every 3 hr during the period. Mean daily cloudiness plotted against daily incident global radiation (Figure 10) shows that, even under full overcast conditions, large amounts of solar radiation were

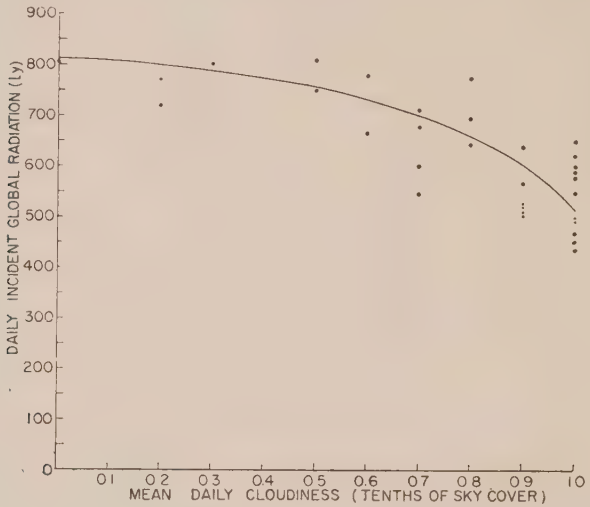

Fig. 10 — Daily incident global radiation vs. mean daily cloudiness, Greenland Ice Cap, 6800 ft msl.

TABLE III

Incident Solar Radiation as a Function of Cloudiness

Mean cloudiness	Radiation received on a horizontal surface, in % of clear day radiation
0	100
5/10	94
10/10	65

received at the snow surface. As shown in Table III, a mean cloud cover of 0.5 reduced the incident global radiation by only 6% and, even under conditions of full overcast, the snow surface received 65% as much global radiation as on a clear day. These high values for measured global radiation with overcast skies over a large snow field have been observed in the Antarctic also (Liljequist, 1953) and a relationship between cloud cover and global radiation similar to that shown in Figure 10 was obtained by Wallen (1949) from measurements made on the Karsa Glacier in Sweden.

Kimball (1919) states that, under fully overcast conditions, there is an average decrease of 70% in global radiation received at the earth's surface in the temperate zone. On the Ice Cap, the decrease was only 35%. The difference may be due to less dense clouds in the Arctic and to multiple reflection between the snow surface and the cloud base.

Angstrom (1934) has estimated the increase of incoming radiation due to multiple reflection from snow covered ground to be about 20% with a clear sky and up to 100% with an overcast sky. According to Angstrom and Tryselius (1934):

$$I_{rd} = I \frac{1}{1 - rd} \tag{5}$$

where:

I_{rd} = radiation income with multiple reflection
I = radiation income for ground absorbing all incoming radiation
r = reflecting power of the ground (0.8 for snow at Ice Cap site)
d = reflecting power of the sky (0.25 for clear sky, 0.75 for overcast sky)

$$\text{with clear sky } I^{\circ}_{rd} = 1.25\, I_{o} \tag{6}$$

$$\text{with overcast sky } I^{10}_{rd} = 2.50\, I_{10} \tag{7}$$

Then:

$$\frac{I_{10}}{I_{o}} = \frac{1.25 I^{10}_{rd}}{2.50\, I^{\circ}_{rd}} = 0.5\, \frac{I^{10}_{rd}}{I^{\circ}_{rd}} \tag{8}$$

Values of I^{10}_{rd} and I°_{rd} for selected periods during 1955 and 1956* are listed below:

Date	I^{10}_{rd}	I°_{rd}	$I^{10}_{rd}/I^{\circ}_{rd}$
July 15-30, 1955	482	804	0.60
July 1-15, 1956	584	892	0.66
August 1-15, 1956	395	583	0.68
Mean			0.65

Values of I_{10}/I_{o} as computed from equation (8) using the above values of I^{10}_{rd} and I°_{rd} are shown below:

Date	I_{10}/I_{o}
July 15-30,1955	0.30
July 1-15, 1956	0.33
August 1-15, 1956	0.34

Similar values for I_{10}/I_{o} were obtained by Wallen (1948).

The value of I_{10}/I_{o} for stations in temperate latitudes has been found to be 0.22 (Kimball, Mon. Wea. Rev., April 1927, Vol. 55). The higher values of I_{10}/I_{o} obtained at the Ice Cap site and by other investigators (Wallen, 1948), (Olsson, 1936), (Mosby, 1932) has been attributed to the arctic clouds being thinner than those in the temperate latitudes.

* Data obtained during 1956 is included in this section.

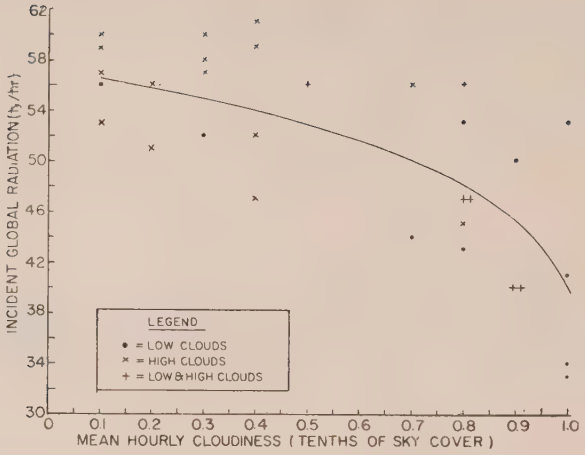

Fig. 11 — Effect of clouds on incident global radiation, Greenland Ice Cap, 6800 ft msl. 0930, 1230 observations.

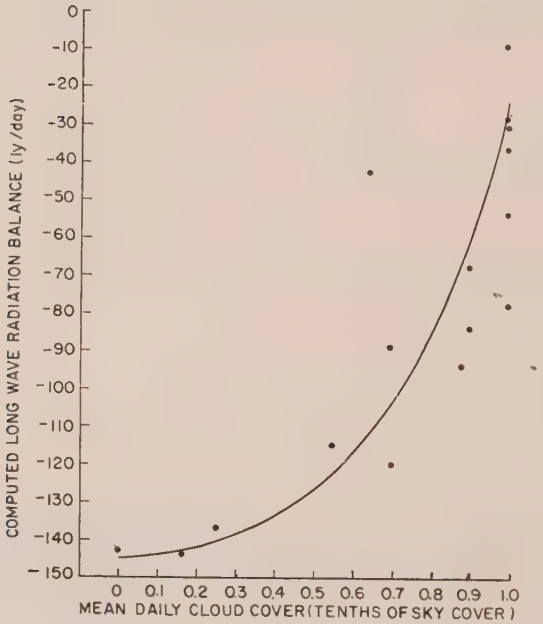

Fig. 12 — Computed long-wave radiation balance vs. cloud cover, Greenland Ice Cap, 6800 ft msl.

The relationship between global radiation received at the snow surface, the extent of cloud cover. and the elevation of the clouds is shown in Figure 11. It appears that low clouds were associated with a more complete cloud cover at this site. Although global radiation was reduced appreciably with extensive low cloud cover, it was still in excess of two thirds of the amount received under clear sky conditions. The diffusion and reflection of large amounts of the incoming global radiation by an extensive cover of low clouds is the primary physical cause of that type of white-out on the Ice Cap in which depth perception is greatly reduced without a corresponding reduction in horizontal visibility. Some studies on this phenomenon are being incorporated in a separate report.

As cloud cover increases, there is a large reduction in the heat loss from the snow surface by outgoing long-wave radiation. In Figure 12, the long-wave radiation balance, as computed from the measurements of net all-wave radiation and net global radiation, is plotted against mean daily cloud cover. Figure 13 shows the observed mean decrease in effective outgoing radiation due to cloudiness, which

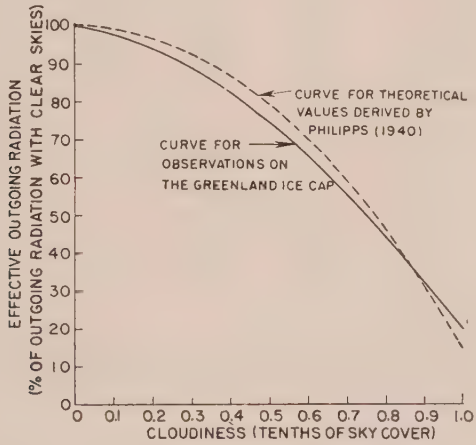

Fig. 13 — Effective outgoing radiation vs. cloud cover, Greenland Ice Cap, 6800 ft msl.

compares favorably with the theoretical values derived by H. Philipps (1940). Since there is only a small decrease in incoming solar radiation with increasing cloud cover, the result is an increase in the net radiation balance (Figure 16).

The net increase in all-wave radiation flux at the snow surface with increasing cloudiness indicates that, with similar atmospheric conditions, ablation would be greater at low elevations and would extend to higher elevations during cloudy summers than during clear summers.

Atmospheric and terrestrial radiation

Incoming and outgoing long-wave radiation was not measured directly, but was computed from measurements made with the net radiometer and the incident and reflected pyrheliometers. The long-wave radiation balance was computed from:

$$I_l - O_l = N - (I_s - O_s) \tag{9}$$

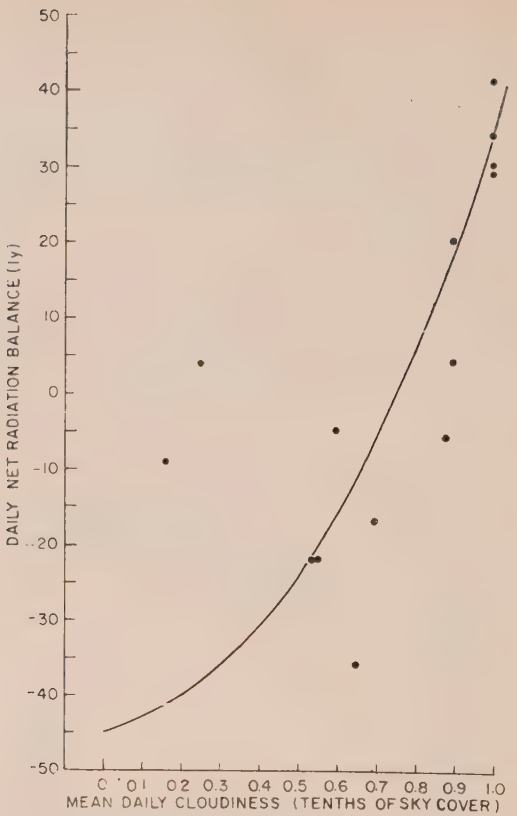

Fig. 14 — Net radiation balance vs. cloud cover, Greenland Ice Cap, 6800 ft ms

where:

$I_l - O_l$ = Long-wave radiation balance or effective outgoing radiation
I_l = Incoming long-wave radiation from atmosphereic water vapor and/or clouds
O_l = Outgoing long-wave radiation from snow surface
N = Net radiation balance
I_s = Incoming solar radiation
O_s = Reflected solar radiation
$I_s - O_s$ = Solar radiation balance

The computed daily values are plotted in Figure 15. The daily balance for long-wave radiation on the Ice Cap, as shown in this figure, is always negative.

Effect of low sun angle on measured radiation

At low angles of incident solar radiation, the response of hemispherically mounted pyrheliometers tends to diverge from the cosine law. Most solar radiation received

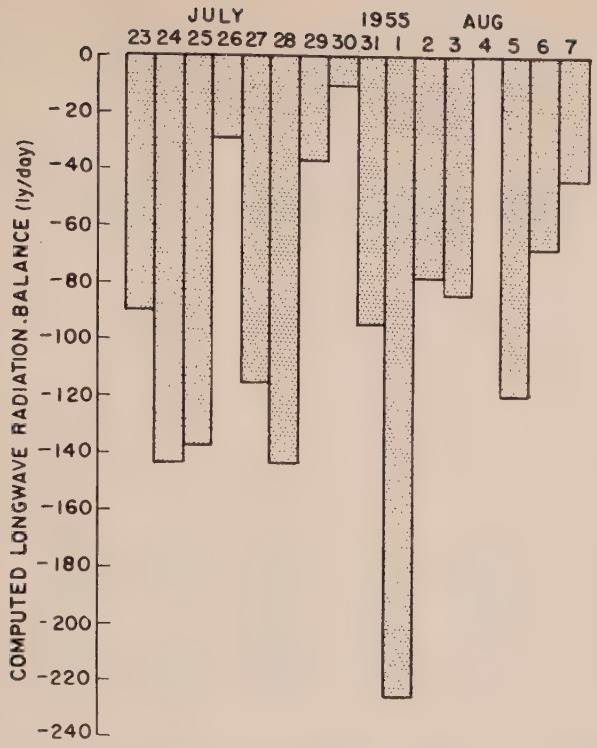

Fig. 15 — Computed daily long-wave radiation balance, Greenland Ice Cap. 6800 ft msl.

in the polar regions during the summer comes from the sun at a very low angle. At the Ice Cap site where these observations were made, the sun angle varied from a high of only 32° at solar noon in July, when these studies were started, to 6° at solar midnight in early August, when the observations were discontinued. Any appreciable divergence of the pyrheliometer from the theoretical cosine response at low sun angles would result in a large acccumulative error in measured global radiation. The instrumentation and procedures used in these studies on radiation permitted evaluation of the effect of low sun angle on the response of the pyrheliometers.

From the equation:

$$I_h' = I_n \sin \phi \tag{10}$$

where:

I_h' = Solar radiation received on a horizontal surface and

I_n = Solar radiation received on a normal surface, the output of the normal incident pyrheliometer was converted to values for solar radiation receivable on a horizontal surface. These values were compared with the values obtained at the same time from the horizontally mounted hemispherical pyrheliometer. The direct solar radiation measured by the hemispherical pyrheliometer was obtained by subtracting the value for diffused sky radiation from the total measured global radiation. Diffuse

sky radiation was measured directly by occulting the sun for brief periods so the hemispherical pyrheliometer received no direct solar radiation.

The data obtained between 1200 and 2400 on one clear day are presented in Table IV and plotted in Figure 16. A mean positive error of 4% in measured global radiation is indicated for the period of the study.

TABLE IV

Effect of Sun Angle on Measured Incident Radiation

Approximate Solar Time	Φ Sun Angle		Sin Φ	I_u Measured	I_h Measured	I_h' Computed
	Deg.	Min.				
2400	6	2	.1045	0.85	0.08	0.09
2300	7	11	.1248	0.92	0.10	0.11
2200	9	3	.1564	0.98	0.14	0.15
2100	11	34	.1994	1.09	0.22	0.22
2000	14	33	.2504	1.17	0.31	0.29
1900	17	50	.3062	1.24	0.41	0.38
1800	21	12	.3611	1.30	0.50	0.47
1700	24	27	.4147	1.31	0.57	0.54
1600	27	20	.4592	1.35	0.65	0.62
1400	31	12	.5175	1.41	0.76	0.73
1200	31	29	.5225	1.44	0.78	0.75
				Total	4.52	4.35
				Mean	0.411	0.395
			Difference between means			0.016
				error		4.0%

For the 24-hr period within which these measurements were made, the measured total global radiation was 804 ly. The apparent error in the pyrheliometer record indicates that this measured global radiation was 32 ly higher than actually received on a horizontal surface. Within this apparent error attributed to divergence from a true cosine response, approximately 8 ly may be the product of temperature changes during the period, leaving a cosine response error for clear-day global radiation of about —3%.

It was not possible to determine whether a similar error might be present in the measurements of reflected radiation. It was observed that, at a very low sun angle, the inverted pyrheliometer appeared to respond to internal reflections of direct solar radiation on the glass globe. This observation is supported by the similar figures for incident and reflected radiation measured during the night hours (Tables 1 and 2, Appendix). Since there were only 4 clear days, and average cloudiness exceeded

0.5 during the period of these investigations, and since albedo measurements appeared to include a similar error, no corrections were applied to the radiation data analyzed in other sections of this report.

Atmospheric transmission coefficient

In passing through the earth's atmosphere, solar radiation is decreased in intensity by scattering due to air molecules and dust particles and absorption by water vapor. The solar radiation which finally reaches the earth's surface may be computed from:

$$I_o = I_o' a^m \qquad (11)$$

where:

I_o = Total solar and sky radiation received on a horizontal surface at the ground, ly/day

I_o' — Total solar radiation received on a horizontal surface at the exterior of the atmosphere, ly/day

a = Atmospheric transmission coefficient

m = Solar air mass (ratio of length of actual path of solar beam to the path through the zenith)

During the period of measurement, a cloudless day occurred on 24 July 1955. From the Smithsonian Meteorological Tables, Table 132 (1951), I_o for that day was computed to be 890 ly and m for the Ice Cap site was computed to be 2.785. The total measured global radiation on that day was 804 ly. Substituting these values in equation (7), the atmospheric transmission coefficient was computed to be 0.964. The high value for the atmospheric transmission coefficient may be attributed either to low atmospheric turbidity and/or the increase in incoming global radiation due to multiple reflection between the snow and sky.

Equation (6) indicates that because of multiple reflection the measured global radiation on a clear day at the Ice Cap site is 25% greater than that measured over a non-reflecting surface. The use of I_{rd} in computing the atmospheric transmission coefficient (a) would result in larger values of (a) than that obtained when the global radiation is not increased by multiple reflection. However, a true value of a may be obtained by using I_o rather than I_{rd}. From equation (6) I_o is equal to 643 ly/day when I_{rd} is 804 ly/day. Substituting in equation (11) a is computed to be 0.888. A value of 0.836 was obtained by Mosby (1932) during the Maud expedition.

Heat balance of snow cover

The upper layers of a snow cover receive heat: (1) from the incoming global and long-wave radiation; (2) by conduction from the air if the air temperature increases with increase of height; (3) by heat of condensation if the vapor content of the air increases with increase of height; (4) by conduction from below if the temperature of the snow increases with increase of depth; and (5) by transfer of heat from precipitation which is warmer than the snow cover.

The upper snow layers lose heat: (1) by outgoing long-wave radiation; (2) by conduction to the air if the air temperature decreases with increase of height; and (3) by conduction downwards if the snow temperature decreases with increase of depth.

Any surplus of heat which the upper snow layers may receive may be used for melting or evaporation, or it may be conducted to lower layers if the snow temperature decreases with depth. For a column of snow in which melting does not occur, a change

in the cold content (defined as the amount of heat required to raise a column of snow 1 cm^2 in cross in section to 0°C) may be expressed as follows:

$$\Delta Q = I + R + Q_s + Q_a + F \qquad (12)$$

where:

ΔQ = Change in cold content, ly (positive for heat gain, negative for heat loss).
Computed from:

$$Q = m\,t\,c \qquad (12a)$$

where:

m = mass of snow, g
t = temperature of snow, °C
c = specific heat of snow, 0.5 cal/g—°C
I = Incident global radiation — reflected global radiation
R = Long wave-radiation balance (positive when incoming and negative when outgoing)
Q_s = Amount of heat received from or lost to the snow, calculated from :

$$Q_s = K\,dt/dz \qquad (12b)$$

where:

K = .0085 ρ^2 (ρ = density of snow)
dt/dz = temperature gradient in the snow
Q_a = Amount of heat received from or lost to air
F = 680 M (latent heat of fusion x mass of sublimate, positive for condensation on the snow, negative for sublimation to the air).

The effective change in the heat balance of the snow cover at the Ice Cap research site was computed from the snow cover studies and meteorological measurements made between 25 July — 6 August inclusive, during which the most complete radiation measurements and snow profile records were obtained.

From snow profile measurements made on 25 July and again on 6 August, the change in cold content of the upper 70 cm of the snow cover was found to be —2 ly, which means that this portion of the snow cover gained 2 ly. Substitution of measured values for I, R, and Q_s in equation (12) permitted computation of the combined valeus of Q_a + F. The results are shown in Table V.

TABLE V

Heat Balance of the Snow Cover at Site 2 for 13 day Period
25 July — 6 August 1956

Gain (ly)		Loss (ly)	Balance (ly)
Global Radiation (I) 1318		Long Wave Radiation (R) 1195	$Q + Q_s + Q_a + F$ + 123
Distribution of heat gain Eq. (8)			
Heat gain, top 70 cm snow (Q)	2 ly		
Heat gain, below 70 cm snow (Q_s)	96 ly		
Total gain for 13-day period, ($Q + Q_s$)			98
Loss by transfer to air and by sublimation to air for 13-day period (Q_a + F)			25

Computation of separate values for Q*a* and F was not possible because of the inability to obtain a precise measurement of the temperature of the snow surface. The mean temperature and dewpoint values for different elevations above the snow surface (Table VI) show that there was a diurnal change in both the air temperature and the dewpoint. The increase in dewpoint accompanying the mid-day rise in air temperature indicates that sublimation from the snow cover to the air took place and probably accounts for most of the computed 25 ly loss from the snow cover by transfer to the air. The 123 ly potentially available to the snow cover was reduced by heat transfer to the air to a net gain of 98 ly for the 13-day period.

TABLE VI

Mean Temperature and Dewpoint at Different Levels
25 July to 6 August 1955.

	Height above Surface (m)									
	0.3	1.2	1.9	3.3	5.5	0.5	1.4	2.1	3.5	5.7
Time	Temperature (—°C)					Dewpoint (—°C)				
0100-0400	10.6	10.8	10.7	10.6	10.6	11.5	11.6	11.8	11.5	11.4
0500-0800	8.4	9.0	8.5	8.6	8.8	11.0	11.3	11.5	11.2	11.2
0900-1200	6.5	6.7	6.2	6.1	6.4	8.9	9.2	9.7	9.0	9.2
1300-1600	6.4	6.7	6.7	6.4	6.5	8.3	8.7	8.9	8.4	8.7
1700-2000	8.3	8.4	8.6	8.3	8.5	9.3	9.4	9.6	9.3	9.3
2100-2400	9.6	9.8	9.8	9.7	9.8	10.1	10.1	10.2	10.0	10.0

Using the seasonal duration of sunshine at 78° N. as a basis for computation, the 98 ly gain by the snow cover over 13 days extrapolates to a 1000 ly gain by the snow cover for the entire 260-day period when solar radiation may be available. This heat gain is sufficient to produce an integrated temperature rise of about 4.5°C in the snow profile extending down to the constant temperature isotherm of —24°C near 30 ft below the surface. Temperature measurements made by SIPRE parties in 1954 in pits at this site showed a 540 ly heat gain by the snow for a 56-day period between 28 June and 22 August, slightly more than half the amount computed as potential gain for the 260 days of possible sunshine in 1955.

For the 13-day period in 1955, there was a net heat balance of 7.6 ly/day. From the deep pit studies made in 1954, the average daily heat balance for a 56-day period was found to be 9.6 ly. The difference, although extremely small, may indicate an actual difference in the summer climate between the two years.

The extremely small supply of heat available to the central ice cap in Greenland is further emphasized when compared with the mean daily heat balance for a snow pack in the Sierra Nevada of California. Table VII shows comparative data for an altitude of about 7000 ft at each research site and for approximately equal periods, although on opposite sides of the summer solstice. No uniform cover of snow exists at the Central Sierra Snow Laboratory after 1 June, and records from the ice-cap site were not available until 5 July.

TABLE VII

Comparison of Mean Daily Heat Balance for a High Mountain Snow Pack in California and the High Ice Cap in Greenland.

Heat Balance	Central Sierra Snow Laboratory California May-June 1952 (ly/day)	Site 2 Greenland July-August 1955 (ly/day)
Radiation, global and long-wave	+ 291	+ 9.5
Exchange between snow and air	+ 109	— 1.9
Net daily gain	+ 400	+ 7.6

The difference between the net daily gain of 400 ly by the snow pack in the High Sierra and the 7.6 ly gain by the central Greenland Ice Cap during the period of approximate maximum radiational heat supply provides one explanation for the enormous climatological difference between two localities where snow is a major factor in the environment. In spite of an annual snowfall equivalent to 80 to 120 in. of water, the entire 10 to 20-ft. deep snow pack which accumulates each winter in the High Sierra is melted by early June. At the same elevation on the Ice Cap, the 5 to 10 water-equivalent inches of snow which fall each year remain to nourish the Ice Cap because of the unfavorable heat balance. The positive heat balance for an entire summer season on the Ice Cap is no more than that accumulated in 2 ½ days in the Sierra. The apparent 2 ly difference in the daily heat budget between the 1954 and 1955 seasons on the Ice Cap is of negligible importance when compared with the 400 ly daily balance available to the high-mountain snow pack in the temperate zone.

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Appendix: Radiation Measurements on the Greenland ice cap, 68 July to August, 1955

TABLE 1 — *Incident Solar Radiation (ly/hr).*

	Date 1955	Time (hr)																							Total
		1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	
July	6	8	11	13	17	23	28	37	54*	62	60	63	62	61	54	47	44	29	22	14	11*	6*	6	5	5	742
	7	7	8*	13	17	19	25	35	36	40	45	43	43	40	38*	33	32	29	20	13	13	10	10	8	9	586
	8	8	12	21	24	26	29	36	34	34	53	48	45	45	42	29	33	24	19	16	12	9	7	6	6	618
	9	11	16	16	28	26	30	36	37	37	46	44	44	37	35	31	29	30	19	14	10	8	8	8	8	646
	10	8	9	14	24	25	28	35	47	44	43	43	42	42	44	44	44	38	32	26	23	10	12	7	7	691
	11	7	11	13	18	22	25	34	38	46	50	63	61	43	37	32	28	25	20	16	12	12	13	7	7	640
	12	7	10	12	16	22	27	38	53	58	60	62	62	59	56	50	45	38	32	26	20	16	12	11	12	804
	13	11	10	13	17	20	27	28	32	36	41	41	40	37	34	30	25	21	16	13	10	6	5	5	5	523
	14	7	11	13	21	33	27	35	55	66	56	50	50*	47	45	39	34	31	25	21	12	7	10	7	6	708
	15	6	6	13	15	20	26	29	39	36	33	36	38	33	39	31	27	22	15	14	11	10	6	5	5	515
	16	8	10	18	17	19	29	31	45	38	38	44	49	43	36	32	29	23	20	13	10	7	6	5	5	575
	17	6	9	11	19	19	25	26	36	40	57	58	44	47	38	36	31	25	23	25	25	14	8	7	7	636
	18	8	10	13	19	28	27	38	51	57	56	63	62	58	55	49	44	35	31	26	12	9	9	6	5	771
	19	7	11	21	15	22	28	35	36	39	52	45	41	50	40	34	26	23	18	13	10	7	5	5	11	594
	20	12	14	23	26	31	38	47	51	51	59	61	59	56	50	48	42	31	26	21	16	13	10	8	7	800
	21	10	13	18	21	29	34	44	49	53	57	38	53	53	53	48	38	33	26	22	16	12	10	9	9	748
	22	10	13	14	15	20	24	30	33	34	38	39	40	39	32	31	26	21	15	11	8	5	4	3	3	508
	23	4	7	9	13	18	25	34	40	44	53	60	59	57	53	48	38	26	18	19	16	11	7	5	8	672
	24	11	15	21	25	30	37	44	50	55	58	60	59	56	53	48	42	36	27	22	15	12	10	9	9	804
	25	11	14	19	24	30	37	44	50	55	58	59	58	56	52	47	41	36	36	11	9	6	5	4	7	769
	26	4	10	10	14	18	25	28	35	37	43	41	45	43	40	39	30	23	16	12	9	7	6	4	6	545
	27	6	10	11	19	20	37	32	36	38	40	49	59	47	44	46	40	34	26	20	15	11	8	7	8	663
	28	9	13	17	22	28	35	41	47	52	55	56	56	53	50	46	40	33	25	17	6	5	5	4	4	719
	29	4	5	7	9	19	26	27	37	38	40	37	36	34	31	31	24	16	15	10	6	5	3	2	3	465
	30	4	5	9	12	17	19	25	34	32	32	39	40	34	30	26	22	17	13	9	5	4	3	3	3	437
	31	4	6	9	11	17	26	29	38	38	47	52	46	44	42	43	31	23	18	16	8	5	4	4	4	565
Aug.	1	5	12	20	32	38	44	47	52	53	56	58	56	52	54	48	38	34	26	20	11	7	5	4	4	776
	2	5	8	9	13	17	26	29	33	38	39	45	38	37	33	28	23	21	18	11	9	5	4	3	4	496
	3	4	5	7	10	15	22	26	32	37	40	44	48	52	48	41	29	20	14	10	6	4	3	2	3	522
	4	3	4	6	9	14	17	23	29	34	37	43	38	42	41	25	19	15	11	10	9	5	5	5	4	448
	5	4	6	8	10	17	24	34	43	47	51	49	49	45	36	38	36	25	23	16	11	7	5	8	7	597
	6	3	8	10	10	13	19	25	31	38	40	51	48	47	43	38	26	20	12	7	5	2	2	2	2	502

July	6	7	8	10	13	17	22	26*	33*	38	40	43	43	41	37	32	32	20	17	11	9	6	5	5	5	520
	7	4	6	9	13	16	21	29	30	33	37	36	36	34	32	28	27	25	17	13	11	10	10	8	8	493
	8	7	8	17	20	23	23	29	28	27	41	40	37	36	31	25	28	21	17	13	11	9	7	7	6	511
	9	7	13	12	21	24	26	31	31	31	39	39	38	38	33	29	27	26	28	17	13	10	10	10	10	563
	10	8	9	11	22	24	28	32	38	37	36	36	36	36	36	37	40	34	31	25	22	10	10	7	6	611
	11	6	8	10	15	18	20	28	31	37	41	50	49	36	35	28	25	23	18	15	13	13	12	8	7*	546
	12	7	8	10	13	16	21	29	37	39	42	45	46	46	43	40	37	32	28	25	20	17	14	13	13	641
	13	11	8	9	11	14	20	22	25	28	32	33	33	32	28	25	21	20	16	13	11	7	6	5	5	435
	14	5	7	9	14	24	23	29	38	44	41	44	48	46	42	40	37	32	27	12	10	5	7	6	5	605
	15	5	10	10	11	17	25	29	33	33	31	34	38	30	36	29	26	20	16	13	11	10	7	4	4	482
	16	5	8	17	16	18	25	29	42	37	38	37	40	35	31	27	25	21	17	22	10	8	7	5	5	515
	17	5	6	8	16	16	21	23	29	31	47	48	37	41	32	32	27	23	22	23	23	14	8	7	6	545
	18	6	7	10	14	22	22	30	40	44	46	50	50	48	46	41	37	31	29	22	11	8	9	7	6	636
	19	6	10	20	14	18	26	36	37	32	44	38	36	44	36	30	26	21	18	14	12	10	8	7	11	554
	20	10	16	21	23	26	26	34	40	43	47	49	50	49	47	42	38	31	27	24	20	17	14	11	9	714
	21	10	13	16	19	24	26	33	38	41	45	48	46	46	45	42	33	30	26	24	20	16	14	13	12	680
	22	12	13	14	14	17	21	28	29	31	34	35	36	36	30	28	25	21	16	13	10	8	7	5	5	488
	23	5	7	8	12	16	22	28	34	38	44	50	51	51	48	44	38	24	17	19	16	11	6	5	8	602
	24	10	13	16	20	24	28	33	38	43	47	50	50	50	47	43	40	34	28	23	19	15	13	11	11	706
	25	11	12	16	20	24	27	32	37	41	45	47	48	47	44	40	37	31	29	10	8	7	5	5	5	628
	26	5	9	9	12	16	22	24	30	32	37	36	40	38	37	36	25	19	15	11	10	7	6	5	5	486
	27	5	8	10	16	16	26	26	31	33	34	49	43	38	40	38	34	28	23	19	15	12	10	8	8	570
	28	8	10	14	17	22	25	30	36	39	43	46	46	46	43	38	33	28	22	16	6	4	4	4	4	584
	29	3	4	10	7	18	22	22	29	28	32	32	32	30	28	25	20	13	13	8	7	6	4	4	2	399
	30	3	4	7	11	14	16	22	28	28	28	34	34	29	25	23	19	17	13	10	6	5	4	3	3	386
	31	4	4	7	10	14	23	25	33	32	30	46	43	37	37	40	28	18	14	11	5	2	2	2	1	477
Aug.	1	1	7	13	20	25	23	31	33	39	41	44	44	41	41	38	32	26	22	17	7	4	3	2	1	555
	2	2	5	6	8	13	20	22	25	29	32	36	31	31	27	23	19	16	13	8	7	4	3	2	2	384
	3	2	2	5	7	11	17	20	25	29	32	35	42	45	41	35	22	16	11	8	5	4	2	1	1	418
	4	1	2	4	6	12	15	20	24	28	31	37	31	37	37	21	16	13	10	8	7	4	4	4	2	374
	5	1	2	5	7	11	17	25	34	37	42	40	41	40	30	32	36	20	20	16	10	8	7	7	6	494
	6	2	2	5	6	5	10	11	25	33	34	46	44	44	40	35	25	18	13	9	7	5	4	4	4	431
	7	4	6	7	14	13	18	29	35	38	41	43	45	44	41	37	26	28	17	18	11	5	6	5	5	536
	8	5	11	14	19	19	28	36	34																	166

*Estimated

TABLE 3

Net All-Wave Radiation (ly/hr)

Date 1955	Time (hr)																								Total
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	
July																									
23	0	0	0	0	1.0	1.0	1.9	3.8	4.7	4.7	2.8	1.0	−1.0	−3.8	−5.6	−2.8	0	0	−6.6	−5.6	−2.8	−1.9	−3.8	−3.8	—16.8
24	−3.8	−2.8	−1.9	−1.0	1.0	5.6*	3.8	4.7	4.7	4.7	3.8	1.9	−0.9	−2.8	−5.6	−6.6	−7.5	−7.5	−7.5	−6.6	−5.6	−5.6	−4.7	−4.7	—44.9
25	−3.8	−3.8	−3.8	−1.9	0	6.6*	3.8	5.6	5.6	5.6	4.7	1.9	0	−1.9	−4.7	−6.6	−4.7	0.9	0	0	0.9	0	0	0	+ 4.4
26																2.8	1.9	0.9	0.9	0.9	0	0	0	0.9	+30.0**
27	0	0	0	1.9	2.8	4.7	4.7	3.8	3.8	3.8	3.8	3.8	0.9	−2.8	−4.7	−6.6	−6.6	−6.6	−5.6	−5.6	−5.6	−4.7	−3.8	−3.8	—22.4
28	−2.8	−1.9	−1.9	0	0.9	2.8	4.7	5.6	4.7	4.7	2.8	0.9	−0.9	−2.8	−3.8	−5.6	−6.6	−6.6	−5.6	0	0.9	0.9	0.9	0	— 8.7
29	−2.8	0.9	0.9	0.9	−0.9	1.9	2.8	3.8	2.8	2.8	1.9	1.9	1.9	2.8	1.9	1.9	0.9	0.9	0.9	0	0	0	0	0.9	+29.0
30	0	0.9	0.9	0.9	1.9	1.9	1.9	2.8	2.8	2.8	2.8	4.7	3.8	3.8	2.8	2.8	0.9	0.9	0.9	0	0	0.9	0	0	+41.1
31	−0.9	0	0.9	0.9	0.9	0.9	0	2.8	2.8	0.9	−1.9	−2.8	0.9	−0.9	−3.8	−1.9	0	0.9	0	−0.9	−1.9	−0.9	−0.9	−0.9	— 5.8
Aug.																									
1	−1.9	0	−4.7	−3.8	−0.9	3.8†	3.8†	3.6†	3.8	4.7	3.8	1.9	1.9	0.9	−3.8	−3.8	−4.7	−5.6	−5.6	−0.9	2.8	0	0	0	— 4.7
2	0.9	1.9	0.9	2.8	2.8	1.9	1.9	1.9	1.9	2.8	3.8	2.8	2.8	1.9	0.9	0.9	0.9	0	0	0.9	0.9	0	−0.9	−0.9	+33.7
3	0.9	0.9	0.9	0.9	1.9	2.8	2.8	2.8	3.8	3.8	2.8	0.9	−2.8	−3.8	−4.7	0	1.9	0.9	0.9	0	0	0	0.9	0	+19.5
4	0.9	0.9	0	0.9	2.8	1.9	2.8	0.9	2.8	2.8	2.8	0.9	0.9	1.9	0.9	0.9	0.9	0.9	0	−0.9	−1.9	−1.9	−0.9	−0.9	+20.3
5	0	0.9	1.9	1.9	2.8	2.8	2.8	3.8	3.8	4.7	1.9	0.9	0	−0.9	−1.9	−4.7	−5.6	−6.6	−6.6	−5.6	−5.6	−4.7	−1.9	−0.9	—16.8
6	−1.9	−1.9	−0.9	−0.9	0.9	1.9	1.9	2.8	3.8	4.7	2.8	1.9	−1.9	−3.8	−3.8	−3.8	1.8	0.9	0	0	0	0	−0.9	0	+ 4.5

*Possible increase due to reflection

†Estimated, blowers blocked

**Estimated

VERTICAL DISTRIBUTION OF VELOCITY IN SALMON GLACIER, B. C.

W. H. MATHEWS
Division of Geology University of British Columbia

Six holes were drilled thermally in Salmon Glacier in the first half of 1956 by Granduc Mines Ltd. to establish the elevation of the rock floor over the line of a proposed low-level access tunnel. Arrangements were made through the National Research Council to case and survey one of the holes to establish the velocity at different levels within the ice. The casing of drill hole No. 4, to a depth of 1,625 feet, at which point the floor of the glacier was apparently reached, was completed on May 1-2.

The first survey was undertaken on May 5-7, and a second survey on August 8-10, 94 days after the first.

The glacier at the drilling site is approximately 1 ½ miles wide and up to 2300 feet deep. Here at the firn line some 4 miles from the source and 10 miles from the most distant terminus, the ice flows in a relatively straight valley trending N45E (magnetic). Transit surveys indicate a surface movement near the center of the glacier of as much as 85 feet in 94 days. Displacement of the collar of the drill pipe which lies ½ mile from the southern side of the glacier was 69.5 feet in this same time interval.

Inclinometer surveys were conducted in both May and August with a Parsons inclinometer which photographs at a predetermined time a pendulum and a magnetic compass. Deviations of the drill hole of more than 10 degrees from the vertical, the limit measureable with the pendulums available in May, necessitated the use of acid bottles below the 900 foot level in the May survey. Instrumental dicffiulties below the 1300 foot level in August made it necessary to assume that the direction of movement above this level, nowhere more than 3° away from N37 ½°E (mag.), prevailed at greater depth. These problems plus the fact that the inclination of an initially crooked hole was being measured at finite intervals, 50 feet, has led to some irregularity in the displacement-depth curve and to a very marked scatter in the shear stress (or depth) to strain rate curve. A third survey *, which it is hoped can be made in 1957 after the surface displacement has reached about 300 feet, should markedly reduce this irregularity and scatter.

Calculations show that the 94 day displacement, in a direction N37 ½° ± 3°E (Mag.), decreased very gradually from 69.5 feet at the surface to about 65 feet at a depth of 875 feet, then more rapidly with increasing depth to 60 feet 1000 feet below the surface and to 50 feet 1300 feet down. Displacement at a depth of 1580 feet in the pipe (actually 1560 feet vertically below the ice surface), the lowest part of the pipe still accessible in August, was about 31 feet, and by extrapolation the sliding of the bottom of the glacier over the rock floor may have been as much as 25 feet.

Shear stresses, which increase with depth as the weight of overlying snow and ice increases, have been computed using the formula:

* The proposed 1957 survey proved unsuccessful, with the casing being obstructed, apparentty by ice, near the collar.

$$\text{Shear stress} = g \text{ Sin } \alpha \int_0^y \rho a y$$

where g is the acceleration of gravity, α is the surface slope of 2°00′, and $\int_0^y \rho \, dy$ is the cumulative load per unit area of snow and ice above depth y computed using the same density distribution as did Gerrard, Perutz, and Roch (1952, pp 553-4) down to 140 meters and a constant density of 0.92 below that level.

Rate of shear strain has been computed for 50 and 250 foot intervals throughout the depth of the drill hole. The shear stress-strain rate curve compares closely with the lower curved portion of Glen's laboratory determined relationship (Gerrard et. al., p. 557), much more so than to the curve determined from the one other fully surveyed drillhole, on a small glacier on Jungfrau, Switzerland (Gerrard et. al., 1952).

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The proposed 1957 survey proved unsuccessful, with the casing being obstructed, apparently by ice, near the collar.

REFERENCE

Gerrard, J. A. F., Perutz, M. F., and Roch, A., 1952: Measurement of the velocity distribution along a vertical line through a glacier; Royal Society, London, Proc., ser. A, vol. 213, no. 1115, p. 546-558.

NEUE ERKENNTNISSE UBER DIE BLOCKBEWEGUNG DER GLETSCHER

Mit einer Diagrammtafel

WOLFGANG PILLEWIZER
München

Summary

Distinct from the well-known type of slow-flowing alpine glaciers are the glaciers, the ice of which is not «flowing» but moving downwards with high velocity as a block between lateral mobile zones. There are three reasons of the block-movement of glaciers:

1. *Vast firn-regions*, the ice-masses of which are pushed through narrow outlets (Greenland glacier).
2. *High accumulation of snow* in small firn-regions of high-mountain glaciers (Rakhiot glacier on Nanga Parbat and all advancing glaciers).
3. *Steep and high slopes of the firn-regions.* At Karakoram-glaciers firn-masses and avalanches are tumbling downwards these slopes, creating the general block-movement of the flat tongues.

Die strömende Bewegungsform der Gletscher

Seit in der zweiten Hälfte des 19. Jahrhunderts mit systematischen Messungen an Alpengletschern begonnen wurde[1]), ist der Bewegungstyp dieser Gletscher bekannt: Es handelt sich um die *strömende Bewegungsform*, bei der die Bewegung im Querprofil am Rande = 0 ist, gegen die Mitte zu gleichmäßig parabolisch ansteigt und dort ihren Höhepunkt erreicht. Auf der Diagrammtafel ist solch ein Bewegungsprofil des Rhonegletschers aus dem Jahre 1875 zu Vergleichszwecken angegeben.

Die Blockbewegung von Inlandeis-Gletschern

Daneben war schon seit den Messungen *H. Rinks* von 1852 und anderer dänischer Forscher bekannt, daß sich die großen Ausflüsse des grönländischen Inlandeises ganz anders bewegen: diese mächtigen Eisströme sind wild zerrissen, sie bewegen sich mit bis zu 20 m Geschwindigkeit im Tag, wobei die Geschwindigkeit in schmalen Randzonen sprunghaft anwächst und nahe dem Rand schon ihren Höhepunkt erreicht. Der ganze, oft mehrere km breite Mittelteil des Gletschers bewegt sich mit gleichmäßiger, nur wenig schwankender Geschwindigkeit als Block zwischen den schmalen Randzonen des jähen Geschwindigkeitsanstiegs.

In neuerer Zeit wurde während der Expéditions Polaires Françaises durch *A. Bauer* der Gletscher Eqip Sermia, ein kleinerer Ausfluß des grönländischen Inlandeises, von 1948-1953 genau untersucht.[2]) Er zeigt bei 3 km Breite ganz typische Blockbewegung mit sehr ausgeprägtem, raschen Anstieg der Geschwindigkeit an beiden

(1) Hier sind vor allem die Messungen am Rhonegletscher in der Schweiz und am Hintereisferner in Oesterreich zu nennen: P. L. Mercanton: Vermessungen am Rhonegletscher 1874-1915, *Neue Denkschr. d. Schweiz. Natf. Ges.* 52, 1916.
H. Heß: Der Hintereisferner 1893-1921, *Z. f. Gletscherkde.*, 13, 1923/24.

(2) A. Bauer: Le Glacier de l'Eqe. Glaciologie Groenland, Vol. II. Expéditions Polaires Françaises, IV, Paris 1955.

Gletscherrändern und mit gleichförmig bewegtem Mittelteil von durchschittlich 3 m Tagesgeschwindigkeit. Das von *A. Bauer* veröffentlichte Bewegungsdiagramm auf S. 84 läßt mit großer Deutlichkeit die beiden scharfen Knicke in der Bewegungskurve zwischen dem gleichförmig bewegten Mittelteil und den mobilen Randzonen des Gletschers erkennen. Diese Knicke sind das wichtigste Charakteristikum für das Geschwindigkeitsdiagramm eines Gletschers mit Blockbewegung.

Die Ursache für die Blockbewegung der grönländischen Gletscher, aber auch anderer, aus Inlandeisgebieten kommender Gletscher [3]), ist in deren riesigen Einzugsgebieten zu suchen, in denen sich gewaltige Eismassen sammeln, die dann durch relativ enge Ausgangspforten zum Meer hinaus gepreßt werden. Dabei entwickelt sich infolge der hohen Geschwindigkeit die Blockbewegung, und die Oberfläche des Eisstroms zerreißt durch die Starrheit der Bewegung in unzählige Eistürme, Schollen und Spalten.

Die Blockbewegung von Hochgebirgsgletschern

Im Jahre 1934 untersuchte *R. Finsterwalder*[4]) die Bewegung des Rakhiotgletschers am Nanga Parbat im NW-Himalaja. Mit Hilfe photogrammetrischer Messungen in 4 Querprofilen stellte er fest, daß dieser Gletscher Blockbewegung mit über 2 m Tagesgeschwindigkeit besitzt, wobei die mobilen Randzonen des jähen Geschwindigkeitsanstiegs stellenweise nur 10 m breit sind. Genau so wie an den grönländischen Gletschern ist die Oberfläche des Rakhiotgletschers sehr stark zerrissen und in einzelne Türme und Schollen zerlegt, die Eigenbewegungen im Gesamtverband des blockbewegten Gletschers aufweisen. *R. Finsterwalder* sprach deshalb von einer « Blockschollenbewegung » des Rakhiotgletschers. Damit war der Nachweis erbracht, daß die Blockbewegungsform auch an Hochgebirgsgletschern auftreten kann, und Nachmessungen, die ich 1954 in denselben Profilen des Rakhiotgletschers vornahm[5]), bestätigten dies, die Blockbewegung verlief sogar um 30 % rascher als 1934. Die Ursache für die Blockschollenbewegung des Rakhiotgletschers ist nach *R. Finsterwalder* in den hohen Niederschlägen zu suchen, die in seinem Firngebiet fallen. Sie erreichen 6-8 m Wasserwert im Jahr, wodurch sich im relativ kleinen Einzugsgebiet des Rakhiotgletschers sehr große Eismassen bilden, die durch das enge Rakhiottal blockartig hinausgepreßt werden. Hervorzuheben ist dabei, daß der Rakhiotgletscher 1934 und 1954 im Rückgang begriffen war; der Firnnachschub aus seinem Einzugsgebiet ist also stets so kräftig, daß die Blockbewegung der Gletscherzunge auch während Rückzugsperioden erhalten bleibt.

Die Blockbewegung vorgehender Gletscher

Die Blockbewegung scheint an Hochgebirgsgletschern des alpinen Typs, die normalerweise die strömende Bewegungsform besitzen, ganz allgemein dann aufzutreten, wenn diese Gletscher kräftig vorstoßen. Leider gibt es erst wenig exakte Bewegungsmessungen an vorgehenden Gletschern, eine Folge des allgemeinen Gletscherrückgangs der letzten Jahrzehnte. Berichte von den großen Gletschervorstößen um die Mitte des 19. Jahrhunderts und Bilder vorgehender Gletscher aus

([3]) Soz. B. der Königsgletscher in NW-Spitzbergen, an dem ich 1938 Blockbewegung von über 2 m/Tag feststellen konnte.

([4]) R. Finsterwalder : Die Gletscher des Nanga Parbat, *Z. f. Gletscherkde.* XXV, 1937.

([5]) W. Pillewizer : Der Rakhiotgletscher am Nanga Parbat im Jahr 1954, *Z. f. Gletscherkde. u. Glazialgeologie*, III, 1956.

dieser Zeit zeigen uns aber stets die Gletscherzungen außerordentlich stark zerrissen, ein Hinweis auf die wahrscheinliche Blockbewegung dieser vorstoßenden Gletscher. *Seb. Finsterwalder*[6]) berichtet von der völligen Zerreißung der Zunge des Vernagtferners während des großen Gletschervorstoßes von 1845-1848. Er berechnete außerdem die Geschwindigkeit des vorgehenden Vernagtferners im obersten Querschnitt zu über 300 m/Jahr = ca. 90 cm/Tag, was ein Vielfaches der gemessenen Geschwindigkeit von 12 cm/Tag im Zeitraum 1893-95 bedeutet. Solch eine hohe Geschwindiggeit muß nach unserer heutigen Kenntnis zur Blockbewegung des Vernagtferners geführt haben, was auch durch die Angaben über die Zerreißung der Gletscherzunge von 1845 bestätigt wird.

Im Jahre 1938 konnte ich mittels photogrammetrischer Bewegungsmessungen am vorgehenden Tindbre in Südspitzbergen feststellen[7]), daß dieser kleine Gletscher vom alpinen Typ sich mit der hohen Geschwindigkeit von 1,20 m/Tag blockförmig bewegte, bei gleichzeitig eintretender völliger Zerreißung seiner Oberfläche. Noch 1937 war die Gletscheroberfläche spaltenlos, und der Gletscher dürfte sich damals strömend mit ca. 30 cm/Tagesgeschwindigkeit bewegt haben. Auch hier war also das Vorgehen eines alpinen Gletschers mit Ausbildung der Blockbewegung verbunden. Erst in neuester Zeit ist es mit dem Vorrücken verschiedener Gletscher wieder möglich geworden, Bewegungsstudien an vorgehenden Gletschern auszuführen. In den Jahren 1952 und 1956 nahm *W. Hofmann*[8]) photogrammetrische Bewegungsmessungen am vorgehenden Nisqually-Gletscher (Mt. Rainier, U.S.A.) vor. Dabei stellte er fest, daß die eigentlich vorgehende Komponente des Nisqually-Gletschers, der Wilson-Gletscher, 1952 unmittelbar am westlichen Eisrand eine Tagesgeschwindigkeit von über 17 cm aufwies; der Gletscher schob sich mit dieser Geschwindigkeit an der Felswand entlang und besaß also typische Blockbewegung, die im Diagramm durch einen senkrechten Geschwindigkeitsanstieg und den Knick in der Bewegungskurve gekennzeichnet wird. Wie weiter unten ausgeführt wird, zeigten die vorgehenden Gletscher des Karakorumgebirges, die 1954 beobachtet wurden, extreme Blockbewegung. Es scheint also ein kräftiger Gletschervorstoß stets mit der Ausbildung der Blockbewegung verbunden zu sein. Die Ursachen liegen, ähnlich wie am Rakhiotgletscher des Nanga Parbat, in einem erhöhten Firnnachschub, der die Eismassen der Gletscherzungen unter verstärktem Druck und entsprechend erhöhter Geschwindigkeit in die engen Hochgebirgstäler preßt, wobei sich aus der strömenden die blockartige Bewegungsform entwickelt.

Die Blockbewegung der Karakorumgletscher

Im Jahre 1954 konnte ich die Bewegung von 6 Gletschern des nordwestlichen Karakorumgebirges mit Hilfe der photogrammetrischen Methode bestimmen. Bei diesen Gletschern handelt es sich entweder um firnfeldlose Lawinengletscher oder um Firnkesselgletscher. In beiden Fällen entspringen flache, schuttbedeckte Gletscherzungen aus gewaltigen Talschlüssen, deren Wände und Steilflanken sich 3000-4000 m hoch erheben. Bei den Lawinengletschern fehlen eigentliche Firngebiete, die Ernährung der tief unter der Schneegrenze liegenden Gletscherzungen erfolgt durch Lawinen, die von den Wänden des Talschlusses abstürzen, und durch einzelne übersteile Hängegletscher, die in Rinnen dieser Wandfluchten liegen. Bei den Firnkesselgletschern

([6]) Seb. Finsterwalder : Der Vernagtferner. Wiss. Erg. H. d. D. u. Oe. A. V. I, Graz 1897.

([7]) W. Pillewizer : Die kartograph. u. gletscherkundl. Ergebnisse der deutschen Spitzbergenexpedition 1938. *Pet. Mitt. Erg. H.* 238, 1939.

([8]) W. Hofmann : Kartographie und Gletscherkunde am Nisqually-Gletscher *Erdkunde* IX, H. 4, 1955, S. 284.

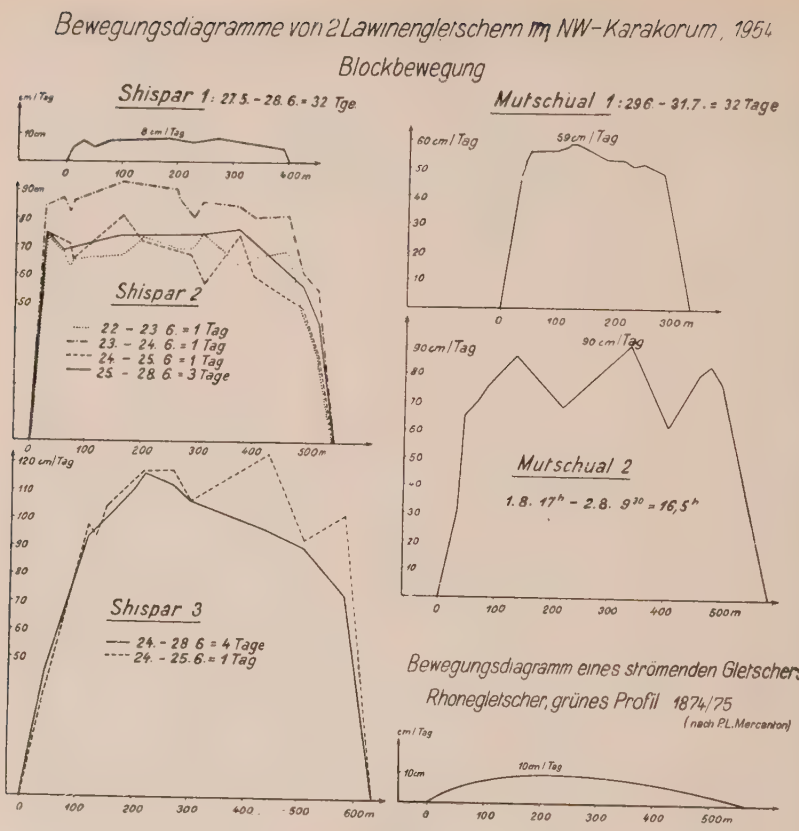

Bewegungsdiagramme von 2 Lawinengletschern im NW-Karakorum, 1954
Blockbewegung
Shispar 1: 27.5. - 28.6. = 32 Tge
Mutschual 1: 29.6. - 31.7. = 32 Tage
Shispar 2
22. - 23.6. = 1 Tag
23. - 24.6. = 1 Tag
24. - 25.6. = 1 Tag
25. - 28.6. = 3 Tage
Mutschual 2
1.8. 17h - 2.8. 9h30 = 16,5h
Shispar 3
24. - 28.6. = 4 Tage
24. - 25.6. = 1 Tag
Bewegungsdiagramm eines strömenden Gletschers:
Rhonegletscher, grünes Profil 1874/75
(nach P.L. Mercanton)

Fig. 1

sind die Steilflanken der Einzugsgebiete firnverkleidet, hier sind also richtige, wenn auch sehr steile Firngebiete, die sog. « Firnkessel » vorhanden.

Die photogrammetrischen Messungen in 21 Querprofilen ergaben, daß alle sechs untersuchten Karakorumgletscher Blockbewegungsform besitzen. Von ihnen ging einer, der Mutschualgletscher, vor, ein Gletscher war stationär und die vier anderen waren im Rückzug begriffen. Auf der beiliegenden Diagrammtafel sind die Bewegungsdiagramme zweier Lawinengletscher der Karakorum-Südseite wiedergegeben. Für das Verständnis der Blockbewegung am wichtigsten sind die drei Profile, die am Shispargletscher gemessen wurden :

Das Profil *Shispar* 1 liegt am Ende der langgestreckten, völlig schuttverhüllten Zunge, wenig oberhalb des eingesunkenen Zungenendes. In diesem Bereich war 1954 die Eisoberfläche wenig zerspalten, während sie im Bereich der Profile 2 und 3 bei gleicher Neigung stark zerrissen war. In Profil 1 betrug der Meßzeitraum 32 Tage; die Geschwindigkeit stieg nicht über 8 cm/Tag, das Profil zeigt aber doch deutlich die Merkmale der Blockbewegung, nämlich den raschen Ansteig der Geschwindigkeit am Rand, die Knicke in der Kurve und die gleichmäßige Bewegung des Mittelteils. Die eingesunkene Zunge und die geringe Bewegung lassen erkennen, daß das Gletscherende 1954 im Rückgang war, es hatte sich auch in den letzten drei Jahrzehnten um mindestens 4,5 km zurückgezogen.

Shispar 2, ein Profil, das etwa 4 km höher den Gletscher quert, zeigt ein völlig anderes Bild; die Tagesgeschwindigkeit stieg hier auf über 90 cm an, die Randzone, in der bereits eine Geschwindigkeit von 84 cm/Tag erreicht wurde, war nur 25 m breit. In diesem Profil wurden die Messungen während eines zeitraums von 6 Tagen viermal wiederholt. Im Diagramm sind diese Wiederholungsmessungen eingetragen. In allen Fällen ist der rasche Geschwindigkeitsanstieg an den Gletscherrändern und der Knick in der Kurve beim Übergang zum Mittelteil zu erkennen. Bei den 1-Tagemessungen zeigen sich starke Unregelmäßigkeiten und von Tag zu Tag bedeutende Schwankungen der Geschwindigkeit, während bei der 3-Tagemessung eine ausgegeglichene Bewegungskurve zustande kommt. In den gegenläufigen Geschwindigkeits schwankungen verschiedener Meßpunkte kommen deutlich die Eigenbewegungen der Schollen und Türme, in die der Eisstrom hier zerlegt ist, zum Ausdruck. Bei längeren Meßzeiträumen gleichen sich diese Eigenbewegungen aus, so daß dann eine gleichmäßige Bewegungskurve entsteht.

Profil *Shispar 3*. Es liegt 2 km oberhalb Profil 2 und konnte innerhalb 4 Tagen zweimal nachgemessen werden. Die Tagesgeschwindigkeit des flachen, stark zerrissenen Gletschers stieg bis auf 1,20 m, die Blockbewegung war deutlich ausgeprägt. Auch hier zeigt die 1-Tagemessung die gleichen Unregelmäßigkeiten wie in Profil 2, bedingt durch Eigenbewegungen einzelner Schollen im Gesamtverband des Gletschers.

Offensichtlich wurde am Shispargletscher in den Profilen 3 und 2 ein Zustand des Vorgehens erfaßt, denn die extreme Blockbewegung dieser Profile läßt sich nur in dieser Weise deuten. Die Welle des Vorrückens war 1954 jedoch noch nicht bis zum Zungenende vorgedrungen, welches noch im Rückzug verharrte. Äußerlich war dies schon am Ausmaß der Zerreißung der Gletscheroberfläche im Bereich der Profile 3 und 2 zu erkennen, während das Zungenende noch unzerspalten war. Trotzdem wies auch dieses in seiner Bewegungskurve die Merkmale der Blockbewegung auf.

Der *Mutschualgletscher* ist der Nachbargletscher des Shispargletschers und war seit 1925 um ca. 7 km zurückgewichen. Im Sommer 1954 war er jedoch wieder im Vorgehen begriffen und zwar schob er sich während des Beobachtungszeitraums von 32 Tagen um 11 m über sein Vorland vor.

Das *Profil Mutschual 1* querte unmittelbar über dem Ende die Wölbung der Zunge und ergab ein Diagramm (siehe Tafel), welches mit Shispar 1 zu vergleichen ist, da dort ebenfalls ein Meßzeitraum von 32 Tagen vorlag. Am Ende des Mutschualgletschers wurde eine Tagesgeschwindigkeit von 60 cm,am Ende des Shispargletschers eine solche von nur 8 cm gemessen. Hier war ein vorgehendes, dort ein zurückweichendes Zungenende vorhanden, beide jedoch mit Blockbewegung, die am Mutschualgletscher besonders typisch und extrem ausgeprägt war. Der flach geneigte, blockbedeckte und am Ende hoch gewölbte Mutschualgletscher war stark zerrissen, von der Wölbung der Zunge stürzten dauernd Eisblöcke auf das Vorland ab.

Das *Profil Mutschual 2* querte 2,5 km höher die flache, in Eistürme und- schollen zerlegte Zunge. Obwohl der Meßzeitraum hier nur 16,5 Stunden betrug, gelang die photogrammetrische Messung dank der hohen Geschwindigkeit des Gletschers einwandfrei. Sie ergab die Bewegungskurve eines blockbewegten Gletschers mit 90 cm Tagesgeschwindigkeit. Typisch für den kurzen Meßzeitraum sind die starken Zacken im Mittelteil der Kurve. Sie deuten ebenso wie bei den 1-Tagemessungen von Shispar 2 und 3 auf Eigenbewegungen einzelner Eisschollen hin, die jedoch immer nur bei kurzfristigen Messungen (1 Tag und weniger) erkennbar sind. Auch im Profil Mutschual 2 hätte ein Meßzeitraum von mehreren Tagen eine ausgeglichene Bewegungskurve ohne die Zacken des Mittelteils ergeben, wie auch die 32-Tagekurve von Profil 1 zeigt.

Die Messungen am Mutschualgletscher haben gezeigt, daß Bewegungsdiagramme von der Art der Profile Shispar 2 und 3 sowie Mutschual 1 und 2 typisch für vorgehende

Gletscher sind. Allerdings muß bei der Konstruktion von Vergleichsdiagrammen darauf geachtet werden, immer denselben Maßstab zu benützen.

Daß ein vorrückender Gletscher Blockbewegung besitzt, ist nicht verwunderlich; die hohe Geschwindigkeit von 1 m/Tag und mehr kann von der strömenden Bewegungsform nicht bewältigt werden, hier muß sich die Blockbewegung entwickeln.

Wie Profil Shispar 1 zeigt, ist im Karakorumgebirge diese Bewegungsform aber auch an zurückweichenden Gletschern mit geringer Geschwindigkeit vorhanden. Das gilt ebenso für die anderen untersuchten Gletscher dieses Gebietes. Sie besitzen Tagesgeschwindigkeiten von 30 bis 60 cm, stets jedoch ist ein jäher Geschwindigkeitsanstieg am Rand und ein deutlicher Knick in der Bewegungskurve am Übergang zum gleichförmig bewegten Mittelteil festzustellen. Sehr charakteristisch sind in dieser Hinsicht die beiden Firnkesselgletscher von Ghulkin und Pasu. Der stationäre Ghulkingletscher erreicht bereits 30 m vom Eisrand entfernt die Geschwindigkeit von 30 cm/Tag, die dann im 600 m breiten Mittelteil erhalten bleibt, und beim 1000m breiten, zurückgehenden Pasugletscher ist die Zone des jähen Geschwindigkeitsanstiegs nur 10 m breit.

Eine Sonderstellung nimmt der 58 km lange *Baturagletscher* ein. Sein mächtiger, 2 km breiter Eisstrom erfüllt ein Längstal auf der Nordseite des Karakorum-Hauptkamms. Die Bewegung wurde in 10 Querprofilen gemessen. Aus den Firnkesseln des Ursprungsgebietes kommt der Baturagletscher mit Blockbewegung von über 50 cm Tagesgeschwindigkeit heraus, wie in den Profilen Nr. 9 und Nr. 8, ca.40 km vom Ende entfernt, gemessen wurde. Im 7 km talab liegenden Profil Nr. 6 ist hingegen kaum mehr etwas von der Blockbewegung zu merken. Hier bewegt sich der Baturagletscher strömend mit 30 cm/Tag, dem Bewegungsdiagramm fehlen die charakteristischen Knicke der Blockbewegung. Unterhalb Profil Nr. 6 mündet in den Baturagletscher ein mächtiger, vom Hauptkamm kommender Firneisstrom ein, der noch 15 km weit als Blankeisstreifen in der sonst schuttbedeckten Gletscherzunge zu verfolgen ist. Durch die Einmündung dieses Seitengletschers erhält der Hauptstrom des Baturagletschers soviel neuen Impuls, daß er sich in den Profilen Nr. 5 (22 km vom Gletscherende entfernt) und Nr. 4 wieder der blockartigen Bewegungsform nähert. Im gleichmäßig bewegten Mittelteil erhöht sich die Geschwindigkeit in diesen Profilen auf 50-60 cm im Tag, die Knicke in den Bewegungskurven sind klar ausgeprägt. Da unterhalb von Profil Nr. 4 keine Seitengletscher mehr einmünden, verflacht die Bewegung in den tieferen Profilen rasch, und in Profil Nr. 1, 4 km vom Gletscherende entfernt, hat der Baturagletscher wieder die strömende Bewegungsform mit nur 12 cm Tagesgeschwindigkeit angenommen, eine Bewegungsform, die er in Profil Nr. 6 bereits hatte. Das schuttbedeckte Zungenende ist schon lange im Rückzug begriffen .

Die Ursache für die Blockbewegung der Karakorumgletscher

Soweit diese Gletscher vorgehen (Mutschual- und oberer Shispargletscher), kann die Ursache für ihre Blockbewegung in einem erhöhten Firnnachschub gesucht werden. Woher kommt dann aber die Blockbewegung der zurückgehenden Karakorumgletscher, die offensichtlich keinen ausreichenden Firnnachschub erhalten? Besonders charakteristisch ist hierfür Profil Nr. 1 am Shispargletscher.

Beobachtungen, die ich 1937 an einigen Gletschern des Jostedalsbre in Südnorwegen machte[9]), können auch auf die Verhältnisse im Karakorumgebirge Licht werfen. An 2 Gletschern des Jostedalsbre-Firnfeldes, am Austerdals- und am Bödalsbre

[9]) W. Pillewizer : Bewegungsstudien an Gletschern des Jostedalsbre in Südnorwegen. *Erdkunde*, V, 1950.

stellte ich fest, daß deren wenig geneigte Zungenenden trotz der geringen Tagesgeschwindigkeit von nur 11-16 cm *Blockbewegung* besaßen. Sie werden aus hohen Eisbrüchen genährt, mit denen diese Gletscher von der Kante der Fjellhochfläche in die tiefen Trogtäler abstürzen. Die Annahme ist berechtigt, daß sich in diesen ca. 1000 m hohen Eisabstürzen die Blockbewegung entwickelt, die dann auch noch in den kurzen flachen Zungen erhalten bleibt, die in den Trogtälern liegen. Sie sind nicht lang genug, um wieder die ihnen eigentlich zukommende strömende Bewegungsform auszubilden.

Im Karakorumgebirge kommt der ganze Firnnachschub für die blockbewegten Gletscherzungen aus den 3000-4000 m hohen Steilflanken der Ursprungsgebiete. Ebenso wie in den hohen Eisbrüchen der norwegischen Gletscher ist auch in den Karakorum-Steilflanken mit einer blockartigen Bewegung des Firns und Eises zu rechnen, denn sie geht über die mehrere 1000 m hohen Abstürze sicherlich so rasch vor sich, daß kein Strömen, sondern eine sich überstürzende Bewegung, also eine Blockbewegung entsteht. Das Extrem bilden dabei die frei stürzenden Lawinen, die ja für die Ernährung vor allem der firnfeldlosen Gletscher von entscheidender Bedeutung sind.

Die langgestreckten, im Talgrund liegenden Gletscherzungen erhalten also den Impuls der Blockbewegung aus ihren Einzugsgebieten und zwar auch dann, wenn kein übermäßig starker Firnnachschub vorhanden ist, denn die Blockbewegung ist hier in erster Linie eine Folge des Steilreliefs und nicht abhängig von der Größe des Firnnachschubs. Auch zurückgehende Gletscher besitzen deshalb die Blockbewegung, die sich erst dann in ein langsames Strömen wandelt, wenn die Gletscherzungen weit genug von ihren blockbewegten Firngebieten talwärts reichen. Nur der Baturagletscher ist lang genug, um den Übergang von der Blockbewegung zur strömenden Bewegung zu ermöglichen. Die Bewegungsmessungen in den Profilen 8-6 und 5-1 des Baturagletschers zeigen, daß für diesen Übergang eine Gletscherlänge von 5-6 km erforderlich ist, auf der aber keine seitlichen Zuflüsse mit Blockbewegung (auch nicht Lawinen) mehr vorkommen dürfen.

Tritt nun zu der Blockbewegung, die durch den Absturz der Firnmassen über die Steilflanken bedingt ist, auch noch eine Erhöhung des Firnnachschubs, so entsteht daraus die extreme Blockbewegung vorgehender Gletscher, und die Gletscherzungen rücken rasch vor. Es braucht hier nicht, wie bei anderen Gletschern, erst die strömende in die blockartige Bewegungsform übergeführt werden, die Blockbewegung ist bereits vorhanden und wird nur verstärkt. Vielleicht liegt hier die Ursache für, das oftmals berichtete sehr rasche Vorrücken von Karakorumgletschern.

Zusammenfassung

Es können drei verschiedene Ursachen für die Ausbildung der Blockbewegung an Gletschern unterschieden werden:

1.) *Riesige Einzugsgebiete*, in denen sich auch bei geringen Niederschlägen große Firnmassen sammeln, die dann als stark zerrissene Gletscher mit hoher Geschwindigkeit durch verhältnismäßig enge Ausggangspforten blockartig hinausgepreßt werden (Gletscher, die von Inlandeisflächen z.B. Grönlands und Spitzbergens ernährt werden).

2.) *Sehr hohe Niederschläge*, die in den Firnregionen von Hochgebirgsgletschern fallen. Hierbei bilden sich ebenfalls große Eismassen, die ihren Abfluß durch enge Hochgebirgstäler finden müssen, wobei sich die Blockbewegung entwickelt (Rakhiotgletscher am Nanga Parbat und ganz allgemein alle kräftig vorgehenden Hochgebirgsgletscher).

3.) *Das Steilrelief des Gletscheruntergrundes.*

Im Karakorumgebirge entsteht die Blockbewegung der flachen Gletscherzungen durch den Absturz von Firnmassen und Lawinen über die hohen Steilflanken der Einzugsgebiete, an manchen Gletschern Norwegens sind hohe Eisbrüche die Ursache für die Ausbildung der Blockbewegung in den darunter liegenden flachen Gletscherzungen. Hier ist also nicht die Größe des Firnnachschubs aus den Ursprungsgebieten, sondern das Steilrelief des Gletscheruntergrundes von Bedeutung für die Entwicklung der Blockbewegung. Sie kann sich leicht extrem verstärken, wenn außerdem noch erhöhter Firnnachschub über die Steilflanken erfolgt, wie es bei vorgehenden Karakorumgletschern der Fall ist.

PHENOMENA ASSOCIATED WITH THE DEFORMATION OF A GLACIER BORE-HOLE

MAYNARD M. MILLER
Lamont Geological Observatory, Columbia University

Abstract

A stress/strain analysis is made of the deformation of a 2-inch (I. D.), 245-foot length of jointed aluminum tubing inserted in a vertical bore-hole in the Taku Glacier of the Juneau Icefield, Alaska. The data embrace three years of consecutive measurement of the changes in azimuth and inclination of the tube in response to the glacier's flow. The results are related to Glen's and Steinemann's exponential flow equations, as derived from laboratory tests on polycrystalline ice, and are compared with bore-hole and tunnel experiments conducted in glaciers of the Alps and Norway. The total depth of ice is known by seismic means. Although the tube penetrates only the upper one-quarter of the Taku Glacier, it is demonstrated that the periodic surveys permit a reasonable extrapolation of the differential flow velocity on a vertical line completely through the glacier. It is shown that with the internal movement dominated by *continuous* plastic creep, a supplemental longitudinal stress trajectory has been superimposed on the normal shear component of flow. In consequence, the low-gradient sectors of the glacier are characterized by a high proportion of basal slippage. This concept is corroborated by the «Block-Schollen» surface velocity profiles and by the existence of tectonic foliation and thrust structures in a narrow zone of *discontinuous* movement near the glacier's margin. Certain anomalies in the bore-hole deformation (creep rate) are explained by configuration of the bedrock floor, and by measured physical differences in the structure of the glacier itself. Observed thermal anomalies on the profile are also considered. These are found to coincide with deformation anomalies and to indicate that zones of inhibited flow occur where colder englacial temperatures prevail.

Introduction

With a sufficiently high stressing condition the internal movement of a glacier can involve processes both of discontinuous shear and continuous flow (Haefeli, 1952). The extent to which these mechanisms co-exist in a large valley glacier has been investigated on the Taku-Llewellyn Glacier system in Southeastern Alaska (Fig. 1). The study has involved determination of profiles of surface and englacial velocity. This has been accomplished by the periodic survey of lines of stakes and other markers on the surface, and by the measurement of changes in inclination and azimuth of an aluminum tube implanted vertically in the glacier. The phenomena associated with the deformation of this tube are discussed in this paper.

Research Site and Techniques

The main research was carried out on a low-gradient portion of the Taku Glacier's névé near Camp 10 (Fig. 1), at a point 3 miles above the mean névé line and 16 miles above the terminus. The drill site (Station 10B, elev. 3600 feet) was on a horizontal movement profile, one mile out from the eastern margin in a sector where the névé was 4 miles wide and where the glacier's depth had been determined by seismic means to be approximately 1000 feet (Fig. 2). A rotary drill rig was used to bore

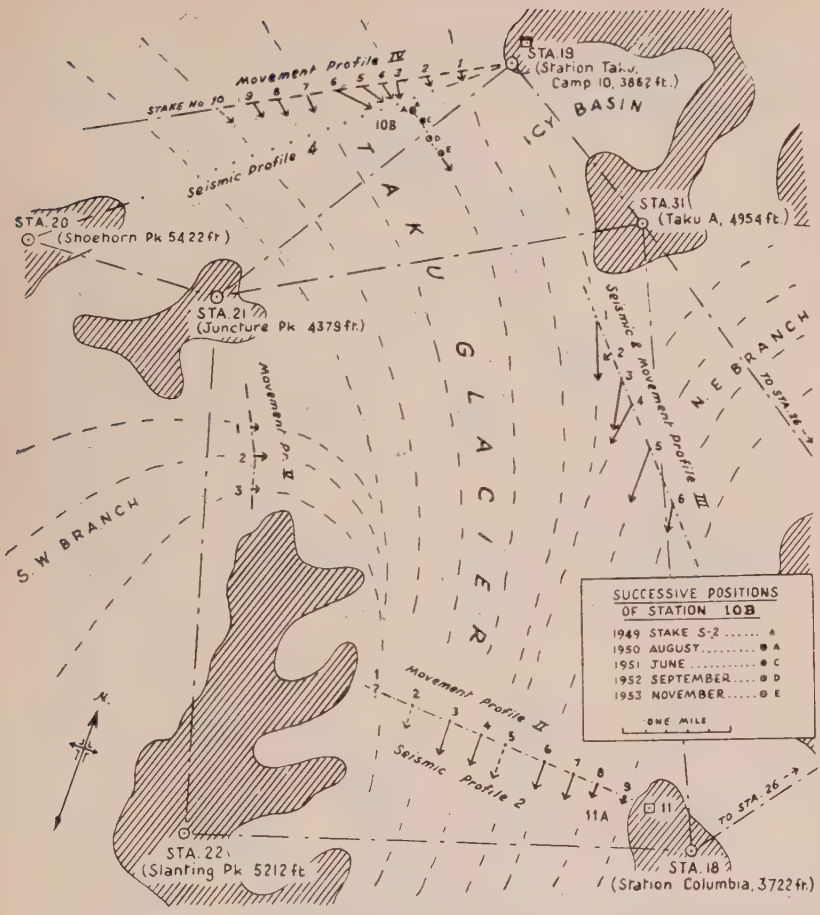
Fig. 1 — Oblique aerial view of Taku Glacier showing tidal terminus in 1929.

a hole 292 feet deep. A 2-inch aluminum pipe, in 10-foot jointed lengths, was inserted into the upper 245 feet of this bore-hole. (1)

Broad and relatively level upland basins comprise most of the accumulation area of this glacier system. Scattered nunataks account for only a small proportion of the surface within the main névé area. Thus there appears to be little opportunity for the ice to be confined or blocked in its movement at depth. A « Block-Schollen» type of flow profile is indicated both on the cross-glacier transect at the drill site (Movement Profile IV) and on another transect at 2600 feet elevation (Movement Profile II), 3 miles below the mean névé line (Fig. 3). On the margins of the glacier in the vicinity of Profile II (e.g. near Site 11A in Fig. 1), a sequence of closely-spaced tectonic foliation structures occurs in the ice, suggesting that the marginal zones of high «mobility» on each side of the glacier may be explained by *discontinuous* laminar movement.

In the central part of each transect, no appreciable discontinuous movement is indicated. This is corroborated in the bore-hole by the fact that in the core samples there was no evidence of over-thrusting. Also in these cores small-scale fracturing, such as that occurring in the marginal zones, appeared to be so restricted that it could account for little, if any, of the mass movement. On empirical grounds, therefore, uniform plastic yielding is considered as the all important mechanism of internal motion in the central part of this glacier. A similar conclusion was reached by the

(1) For this research, a Pioneer Straight-line drill rig was employed, loaned by the E. J. Longyear Co. of Minneapolis, Minn. The drilling equipment and accessories totalled 8 tons and were delivered to the research site by ski-aircraft supporting the Juneau Icefield Research Project (Miller, 1951).



1950 TO 1953 GLACIER SURFACE MOVEMENT SURVEYS IN VICINITY OF CAMPS 10 AND 11

Showing Direction and Relative Magnitude of Average Daily Movement across Key Depth Profiles During Months of July and August. Velocity Scale Exaggerated 500 X. (Nunatak outlines only approximate and base triangulation network diagrammatic)

Fig. 2

Jungfraujoch Research Party (Perutz, 1940, p. 134) on the Aletsch Glacier in Switzerland, an ice stream of similar size and form to the Taku.

Depth of Firn

During the drill operations all core samples were structurally analyzed as they were brought to the surface. A continuous sequence of firn was identified down to 120 feet. A relatively dense material (firn-ice?) was encountered below 85 feet and cores of what appeared to be mostly aerated bubbly ice were recovered between

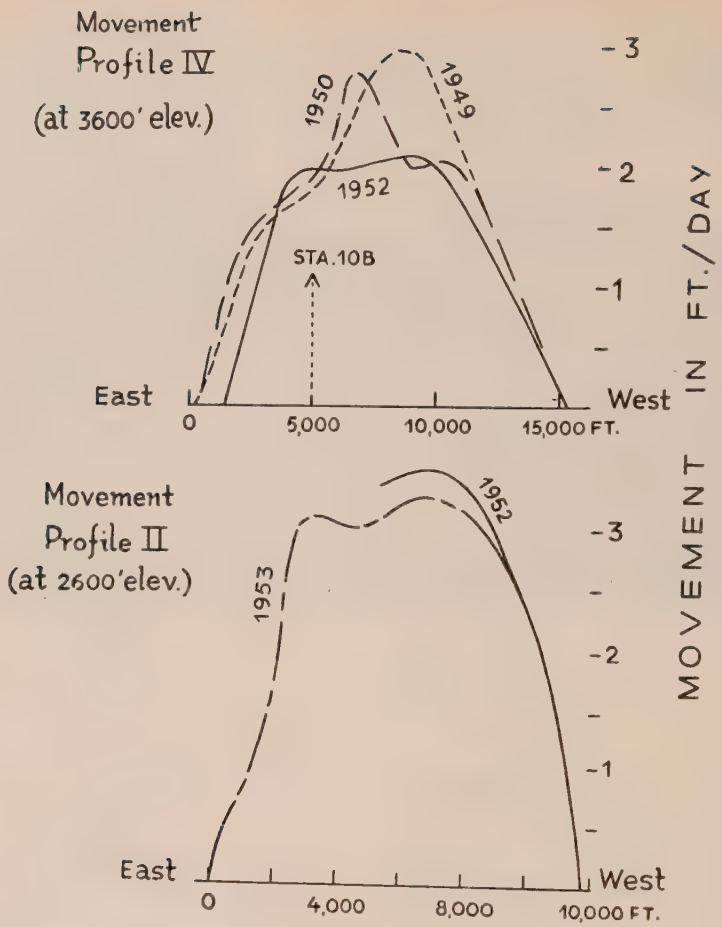


Fig. 3 — Surface velocities in the Névé-Line zone of the Taku Glacier.

120 and 140 feet. In this zone the cores were so crushed and distorted by the churning action of the drill that they were difficult to assess. Below 140 feet, however, only solid ice was encountered. The thickness of firn is thus considered to be about 120 feet, with an unexpectedly abrupt change indicated between the firn-pack and deeper ice at this site (Fig. 4). Evidence from elsewhere on the icefield has suggested that this sudden change is related to a pronounced alteration of climatological conditions during the early decades of the last century (Miller, 1956, ch. XII, B4).

The Bore-hole Surveys

The interior of the aluminum tube was surveyed with a magnetic inclinometer manufactured by the Eastman Oil Well Survey Co. of Denver, Colorado. The shot points were spaced 5- to 20-feet apart along the tube. Pertinent statistics over a

38-month period are noted in Table I. The first survey was in August 1950 and the last in November 1953. At this time, the protruding top of the pipe was found to be excessively tilted down-glacier and to have become constricted by buckling somewhere in the upper 50 feet so that the inclination could not be measured at depth. Details of the surveys, and plan and cross-sectioned plottings of the results, have been given in previous publication (Miller, 1954a).

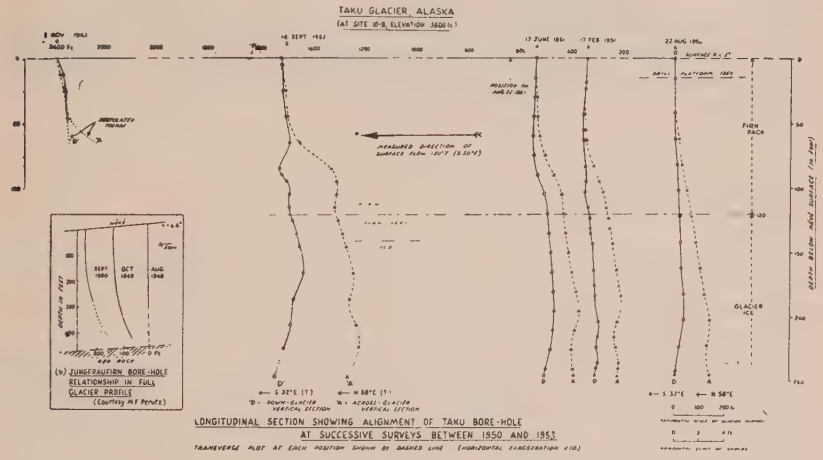

TAKU GLACIER, ALASKA
(AT SITE 10-B, ELEVATION 3600 ft.)
14 SEPT 1952
17 JUNE 1951
17 FEB 1951
22 AUG 1950
MEASURED DIRECTION OF SURFACE FLOW
FIRN PACK
GLACIER ICE
(b) JUNGFRAUFIRN BORE-HOLE RELATIONSHIP IN FULL GLACIER PROFILE (COURTESY M.F. PERUTZ)
LONGITUDINAL SECTION SHOWING ALIGNMENT OF TAKU BORE-HOLE AT SUCCESSIVE SURVEYS BETWEEN 1950 AND 1953
TRANSVERSE PLOT AT EACH POSITION SHOWN BY DASHED LINE (HORIZONTAL EXAGGERATION ×10)

Fig. 4

TABLE I

Depth and Inclination Data on Taku Glacier Bore-hole Surveys

Date and Year	True Vertical Depth of Survey (²) (in ft.)	Time Increment Since Previous Record (months)	Max. Drift Angle to S.W.	Survey Spacing; Remarks
22 Aug., 1950	245	—	1°10′ (at 102 ft.)	20 ft.
11-17 Feb., 1951	244.96	6	2°00′ (at 94 ft.)	5 to 20 ft.
17 June, 1951	239.95	4	3°00′ (at 89 ft.)	15 ft.; protective casing lost in bottom 5 ft. of pipe.
14 Sept., 1952	222.73	15	7°15′ (at 84 ft.)	10 to 20 ft.; only false bottom attained 22 ft. from true bottom of pipe.
2 Nov., 1953	25	13.5	?	12 ft.; constriction in pipe at 25 ft. precluded deeper survey.

(²) Reference plane, 15 feet above wooden drill platform (Fig. 4).

All readings are within the manufacturer's designated accuracy of the inclinometer and have been plotted as true deformation (Fig. 4). This is on the assumption of no differential horizontal or vertical movement between the tubing and its encompassing firn or ice. Also during the three years of accumulation involved, the change in vertical position of the upper end of the pipe with respect to the regime surface was small. The survey depths in this figure are therefore based on the initial reference level as if it were the true glacier surface.

At the time the pipe was installed, the maximum drift angle was 1°05′. When taking into account the down-valley glacier gradient (1° at Site 10B), the bore-hole was essentially perpendicular to the ice surface. This implies that all subsequent deformation of that portion of the pipe resting in solid ice cannot have been significantly affected by the initial alignment.

The Vertical Velocity Distribution

The deformation of the tube is considered to reflect the differential down-glacier movement of this ice mass at depth. A certain amount of vertical strain in the firn (compaction) may have accentuated the normal creep strain in the upper third of the tube, but this effect is considered slight.

Between initial and final surveys, the top of the pipe travelled a map distance of 2250 feet down-glacier in the direction 130°T. The actual course was somewhat irregular, with a total surface movement of 2378 feet. The mean aunual rate was 761 feet with a fairly uniform surface velocity prevailing throughout each year. The net rate of movement, however, was 10 per cent *greater* in winter, presumably due to the increased pressure of new accumulation.

Over short periods, faster rates and changes in direction of surface movement occurred. This is illustrated by the 1952 survey which shows that between July 19th and August 26th the flow direction shifted nearly 40° to a more southerly course. According to the triangulation, which is believed to be correct, the surface velocity increased at the same time to 4.2 feet per day. The cause of this is not clear, but it suggests relationship to an increased measure of extending flow either due to a change of load or to the fact that the glacier passed over a different configuration of the sub-glacial floor. (3)

The increments of relative movement between successive surveys at depth are illustrated in Figure 4. It is clear from this diagram that a somewhat greater flow rate occurs in the surface zone. In general, the vertical velocity distribution is similar to that measured in the upper half of the Jungfraufirn pipe in 1948-50 by Gerrard, Perutz and Roch (1952; v. inset in Fig. 4). Since there is a 1 to 10 horizontal exaggeration in the Taku Glacier curves, it may be seen that the differential velocities are not large. The inclination measurements, however, are more detailed and indicate certain features not found in the Jungfrau profiles.

Although the surface surveys show that the primary direction of mass movement is S.E. (down-glacier), the englacial surveys indicate that there is a marked lateral component of differential adjustment internally, in a cross-glacier direction towards the N.E. This is expressed as a gradual leaning of the pipe, reflecting a tendency of the ice to be diverted into a broad marginal basin (Icy Basin, Fig. 2). Half a mile east

(3) That abnormal «currents» or surges of flow may occur in portions of the glacier is suggested by the 1949 and 1950 velocity curves of Profile IV (Fig. 3). Note the narrow zone of faster moving ice, 2000 to 4000 feet west of Station 10B, a condition apparently not persisting into 1952. As discussed later, it is of interest to consider whether such variations in flow rate bear any relationship to varying glaciothermal conditions.

of the bore-hole, the névé inclines abruptly into this depression so that the sub-glacial floor is also assumed to slope away from the longitudinal trend of the valley. Although at Station 10B the névé is relatively flat, half a mile to the west the gradient inclines in the opposite direction towards the center of the trunk valley. Thus the pipe would appear to have passed along the crest of a low threshold delimiting the western edge of the tributary basin.

In order to distinguish the most significant trend of internal flow, the interplay of stresses resulting from the bedrock slope and configuration is briefly considered. For this purpose the tangent of the pipe's tilt angle has been plotted as a function of depth (Fig. 5). Only those values with drift directions lying within 20° of the

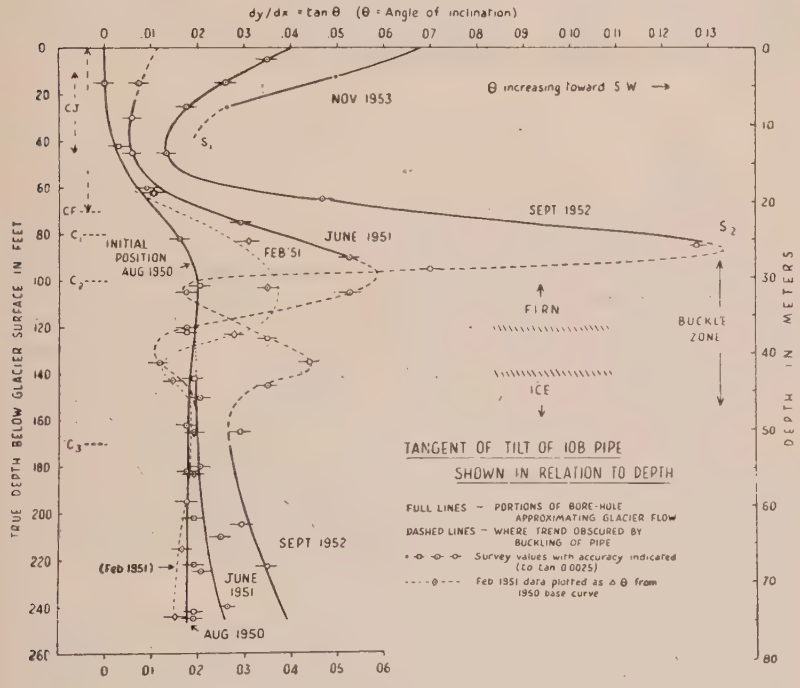

dy/dx = tan θ (θ = Angle of inclination)
0 .01 .02 .03 .04 .05 .06 .07 .08 .09 .010 .011 .012 .013
θ increasing toward S.W →
NOV 1953
SEPT 1952
JUNE 1951
FEB '51
INITIAL POSITION AUG 1950
FIRN
ICE
BUCKLE ZONE
TANGENT OF TILT OF 10B PIPE
SHOWN IN RELATION TO DEPTH
FULL LINES – PORTIONS OF BORE-HOLE APPROXIMATING GLACIER FLOW
DASHED LINES – WHERE TREND OBSCURED BY BUCKLING OF PIPE
Survey values with accuracy indicated (to tan 0.0025)
Feb 1951 data plotted as Δ θ from 1950 base curve
(Feb 1951)
AUG 1950
TRUE DEPTH BELOW GLACIER SURFACE IN FEET
DEPTH IN METERS

Fig. 5

cardinal directions N.E. or S.W. are used. Thus the figure represents the *maximum deformation* from the surface slope involved, and with the changes of inclination taking place in essentially one plane (the tilt angle in succeeding surveys increasing towards the southwest). The accuracy of the surveys is indicated to tan 0.0025, which is equivalent to the maximum percentage error in the inclination surveys. The results show that in this plane the *lowest 100 feet of pipe exhibit a much more uniform tilt than the upper 150 feet*, a fact of considerable importance to the interpretations.

Temperature vs. Deformation Anomalies

The most abrupt change appears as a flexure at 85 feet. It was at this depth that firn cores began to change over to those of firn-ice and bubbly ice. It is concluded

that this flexure is largely the result of a relatively faster creep of firn over the deeper glacial ice, a condition not unlike that observed in seasonal snow-packs on bedrock (Haefeli, 1948). The flow pattern at shallow depth is complex, since a faster relative movement at the surface gradually decreases to a minimum at 45 feet and then rapidly increases again to a maximum at the 85-foot flexure. Such a reversal in flow gradient may be explained by the glaciothermal conditions indicated in Figure 6. Here the base of the annual cold wave in 1950-51 is shown to lie at 65 feet, which is equivalent to the CF level noted on the ordinate of Figure 5. The marked increase in deformation at and below this depth would thus appear to be a correlate of the temperature condition — i.e. with more rapid flow in the underlying or isothermal part of the firn-pack. Glen (1955, pp. 528-9) has shown experimentally the great sensitvity of the creep rate to temperature. For example, according to his painstaking laboratory tests the minimum rate of strain ($\dot{\gamma}$) of ice under constant stress (i.e. 6 kg cm^{-2}) varies with the temperature (t) in the ratio:

t	$\dot{\gamma}$ as a proportion of the value at 0°C.
at — 1.5°C.	13.5%
at — 6.7°C.	4.7%
at — 13.0°C.	1.0%

At an ice temperature of only 1°C. below freezing, the flow rate may be expected to be slowed down to approximately 1/5th that of conditions at 0°C. And at the maximum chill, shown in Figure 6 (—9.5°C.), the internal creep within the zone affected by the annual temperature wave should be less than 1/25th of that under the same stress at 0°C. With these proportions pertaining under conditions of unchanging stress they should be considered only qualitatively. For instance since the heat of deformation cannot escape from the interior of a temperate glacier, it may be generally assumed that there is local melting with a consequent weakening of crystal structure. It follows that the flow behaviour of an isothermal glacier will depend somewhat on its flow history, which in turn means that one field case may not be directly comparable with another. On the other hand, in an experimental thermodynamic treatment of the temperate state of ice, Steinemann (1957, personal communication) has found no reason to refute the simple extrapolation to 0°C. of flow data obtained under subfreezing conditions.

Thus it may be expected that *vagaries of the 10B tilt curve bear a strong relationship to the thermal conditions.* A further review of the pertinent field data appears to bear this out. For instance, not only does the depth of maximum annual penetration of the 1951 cold wave correspond with a pronounced flexure in the pipe, but also the slight temperature depressions noted at 80, 100 and 170 feet in Figure 6 lie at the C_1, C_2 and C_3 flexure points in Figure 5. This suggests that entrapped sub-freezing temperatures may in fact play a part in the differential flow of firn-ice (and ice) at depth. (4) Further to this the zone of delayed chill, which was retained up until the end of spring between the 15- and 50-foot depths (Fig. 6), corresponds in position with the uppermost flexure in Figure 5 and hence lies at the horizon of least velocity change (v. line CJ). *The noteworthy correspondence of these thermal anomalies with the most striking anomalies of deformation suggests not only a direct causal relationship but a similar temperature regime in each of the referenced years.*

Additional flexures occur in the tangent of the tilt plot at the 135-foot and 225-foot depths. Original alignment produced in the drilling may explain the lowermost

(4) The correspondence of position argues against the possibility of these anomalies being due to thermistor error.

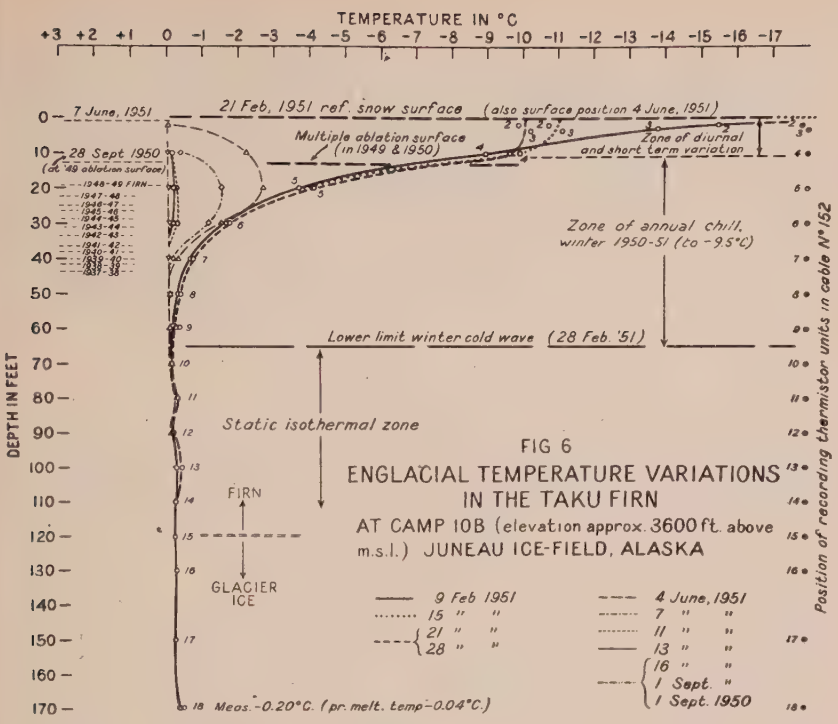

TEMPERATURE IN °C
+3 +2 +1 0 -1 -2 -3 -4 -5 -6 -7 -8 -9 -10 -11 -12 -13 -14 -15 -16 -17
7 June, 1951
21 Feb, 1951 ref. snow surface
(also surface position 4 June, 1951)
Multiple ablation surface
(in 1949 & 1950)
28 Sept 1950
(at '49 ablation surface)
Zone of diurnal
and short term variation
Zone of annual chill,
winter 1950-51 (to -9.5°C)
Lower limit winter cold wave (28 Feb. '51)
Static isothermal zone
FIRN
GLACIER
ICE
DEPTH IN FEET
Position of recording thermistor units in cable N°152
9 Feb 1951
15 " "
21 "
28 "
4 June, 1951
7 " "
11 " "
13 " "
16 " "
1 Sept. "
1 Sept. 1950
Meas. -0.20°C. (pr. melt. temp -0.04°C.)

FIG 6
ENGLACIAL TEMPERATURE VARIATIONS
IN THE TAKU FIRN
AT CAMP 10B (elevation approx. 3600 ft. above m.s.l.) JUNEAU ICE-FIELD, ALASKA

Fig. 6

anomaly, but that at 135 feet is more likely due to excessive bending of the pipe at the transition point between the firn-pack and deeper solid ice. (Hence referred to as the «buckle zone» in Fig. 5). The possibility is minimized, therefore, that either of these flexures bears any genetic relationship to a discontinuity in the thermal profile.

It is of further interest that the dominant direction of surface movement given by the theodolite surveys is 130°T., while the englacial records show the maximum deformation in the direction 150°T. Most of the deformation may have occurred during those shorter periods when the pipe was transported along the more southerly course. However, it would seem more likely that a torsional stress has been set up as a result of the easterly component of firn creep acting on the top of the pipe in one direction while the southerly component of ice flow at depth affected it in another. A further possibility is that this effect is due to a vertical shear strain superimposed on the flow strain as a result of some settling in the firn. The resulting differential movement may have served to twist the top of the pipe in a counter-clockwise manner. Regardless of the actual nature of the stressing conditions, there is a marked difference in the recorded deformation of the lower half of the pipe *imbedded in solid ice compared to that in the upper half imbedded in firn. Therefore, only the data from the lower section are used as a key to the stress/strain conditions in the glacier as a whole.*

Creep Rate as a Function of Load

The data in Figure 6 show that the ice at depth in the 10B profile is essentially temperate — i.e. isothermal at the pressure melting point. The fundamental relationship as far as the creep rate is concerned is, therefore, the dependence of rate of strain

on the load. In the analysis the first step is to resolve the curves of Figure 5 into a figure of strain as a function of the stress. This is done on the assumption that ice behaves as a plastic solid and with the starting consideration that the strain rate may be calculated as a problem in pure shear (Nye, 1951). An admirable study of the creep of polycrystalline ice has recently been made on laboratory samples by Glen (1955) who has shown that the rate of strain is basically the result of the thermal motion of molecules tending to re-arrange themselves in a pattern conditioned by the stress. That these laboratory tests serve at least as a working basis for the analysis of glacier flow has been reasonably demonstrated by the comparison of theoretical results with those achieved in field measurements from bore-holes and tunnels in Alpine and Norwegian glaciers (Nye, 1953). Accordingly, as was first suggested by Perutz (1950), the exponential relationship between the stress and the deformation per unit time is taken as:

$$\frac{d\gamma}{dt} = k\ \tau^n$$

where γ is the shear strain, τ is the shear stress in bars, k is a constant for any given temperature and n is an empirical constant depending in large measure on the physical character of the ice (i.e. its texture, intergrowth complex, grain size, and preferred crystal structure). Also, the exponent n probably depends to some degree on the magnitude of the stress, a factor considered by some to be significant only in very deep or otherwise highly stressed ice. ([5]) In this case $\frac{d\gamma}{dt}$ is considered as the *strain per year*, and is designated as $\dot{\gamma}$. The function is readily calculated as radians of angular deformation per year, since it is numerically equivalent to the tangent of the inclination angle θ in Figure 5.

Although from Steinemann's experiments (1954, p. 449) it must be expected that the power law used with a universal constant n can give no more than a good approximation of the true strain rate at depth (also v. previous footnote), it is probable that because of the simple form and known thicknesses involved in the Taku Glacier fairly representative results can be obtained through a measured mean value of the factor n. By applying the formula, therefore, it is considered that a reasonable determination is made of the movement *within the glacier*. This computation should at

([5]) It has been mentioned that Glen's experiments have demonstrated the rapid increase in flow near 0° C. This is because within the temperature limits 0.02° to —13.0° C. k is largely a function of temperature. For the temperature range from —1.9° to —22° C. and larger stress limits, Steinemann (1956, and personal communication) verified Glen's general conclusions through compression and tension tests on polycrystalline natural ice (melted snow). As Glen has done, he distinguishes a secondary or quasi-viscous flow and a tertiary or paracrystalline (metamorphic) flow. During the latter phase at varying stresses, he observed the structural behaviour of his samples to be unstable. He proposes that the tertiary flow is identical to processes in a glacier—apart from a supplemental stressing which would affect k and probably also the exponent n. *From his considerations, for both stages the flow law would have an exponent n which, depending on the stress, varies from 1.6 to 4 for stresses from 0.7 to 17 kg cm*$^{-2}$. Additionally, Steinemann has evidence that the plasticity of single ice crystals is influenced by the type and magnitude of impurity in the water from which the ice has crystallized (a decrease in purity causing an increase in k). This effect, however, must be vanishingly small in the present field case, since the only significant impurities in firn and ice samples of the 10B bore-hole were chloride and NaCl in the minute proportions of 0.3 to 7.4 p.p.m. (Miller, 1953, pp. 33-36).

In Glen's laboratory experiments, at —0.02° C. and with a load of 1 to 10 kg cm^{-2}, the constants were: $n = 4$ and $k = 0.25$. The values of each constant as used in the Taku Glacier analysis are discussed in the following pages. It will be shown that the value of n in this particular case approximates the mean of the range found by Steinemann.

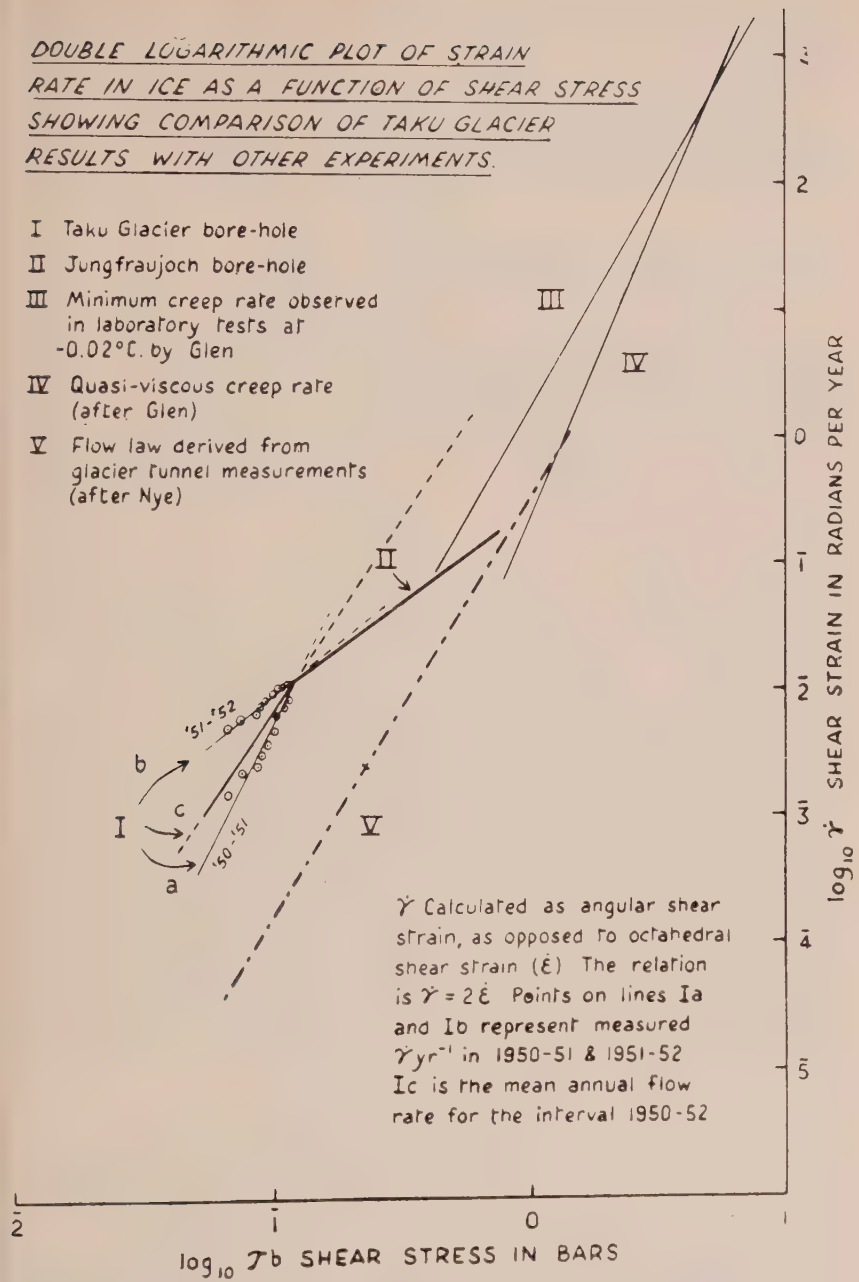

DOUBLE LOGARITHMIC PLOT OF STRAIN RATE IN ICE AS A FUNCTION OF SHEAR STRESS SHOWING COMPARISON OF TAKU GLACIER RESULTS WITH OTHER EXPERIMENTS.
I Taku Glacier bore-hole
II Jungfraujoch bore-hole
III Minimum creep rate observed in laboratory tests at -0.02°C. by Glen
IV Quasi-viscous creep rate (after Glen)
V Flow law derived from glacier tunnel measurements (after Nye)
'51-'52
'50-'51
a
b
c
γ̇ Calculated as angular shear strain, as opposed to octahedral shear strain (ε̇) The relation is γ̇ = 2ε̇ Points on lines Ia and Ib represent measured γ̇yr⁻¹ in 1950-51 & 1951-52 Ic is the mean annual flow rate for the interval 1950-52
log10 γ̇ SHEAR STRAIN IN RADIANS PER YEAR
log10 τb SHEAR STRESS IN BARS

Fig. 7

least be more reliable than for glaciers of complex morphology. Also, by combining the englacial movement data with the surface velocity and geophysically determined depth records, an assessment is made of the relative proportion of movement through *slippage on the bed.*

The shear stresses indicated in the 1950-51 and 1951-52 regime years have been derived from the curves of Figure 5, assuming from the depth profiles that the glacier lies on a gently sloping bed without significant relief. The *shear stress* at each inclinometer level is calculated from the basic formula in dynes cm^{-2}:

$$\tau = D\rho g \sin \alpha$$

Where D is the depth in centimeters, ρ the bulk density of the overlying mass ([6]) α the surface gradient and g the gravitational constant.

The results are shown in Figure 7 (Plot I) as a double logarithmic plot of the strain rate against shear stress. Two sets of curves (I*a* and I*b*) were obtained by taking the difference between the smoothed full-line curves of 1950-51 and 1951-52 in Figure 5, thus giving a comparison between the englacial flow rates for the first two years of record. Since the concern of this paper is the strain relationship in solid ice, the plottings represent only the depths between 120 and 245 feet. ([7]) The values obtained from the actual inclinometer readings are indicated by small circles in the figure to show the close agreement with the smoothed plot. It is of interest that the points in each year's record are all on a straight line and that the shear stresses involved are all below 1 bar. ([8]) As was found by Gerrard, Perutz and Roch in the Jungfraujoch bore-hole (1952, p. 554), there is no indication of a sudden yield stress and the slope of each line is much steeper than would be the case were we dealing with a Newtonian liquid. In fact, by comparing the 1951-52 curve (line I*b*) with the slope calculated by Perutz for the Jungfraujoch bore-hole (line II) we find very close agreement. This is of interest because the results both apply to temperate ice.

It is of further interest to compare the Taku Glacier results with the compression tests carried out by Glen, and with the flow law derived by Nye (1953) from analysis of glacier tunnel measurements. The comparison is made in Figure 7, which shows a double logarithmic plot of the data for each of the other experiments, presented as a log of the strain rate against the log of the shear stress. To facilitate the comparison a plotting of the mean flow rate per year in the Taku Glacier example is also given for the interval 1950 to 1952 (line I*c*). This may be considered as a measured mean creep rate at the bore site because of the uniform increase in surface gradient between these two years. In this figure, $\dot{\gamma}$ is shown as the angular shear strain rate as opposed to octahedral shear strain rate $\dot{\varepsilon}$. ([9]) The relation is $\dot{\gamma} = 2\dot{\varepsilon}$. The minimum creep rate observed in Glen's laboratory tests at —0.02°C. is shown (line III) in comparison with the quasi-viscous creep rate (line IV, also from Glen's experiments) and the flow law relationship after Nye (line V). The mean annual strain rate from the Taku measurements is closely parallel with that of line V, although representing a considerably lower range of stress.

In each of these lines, a different value pertains for the empirical constants n and k. From the cited laboratory test, the quasi-viscous line may be considered as the idealized relation in which the full stressing conditions are known. As such it provides us with an established reference, any deviations from which may be attributed to a difference in the law, or to the presence of extending and compressive

([6]) From graphic integration of field data and extrapolated density curves in the vicinity of the bore-hole.

([7]) An analysis of the strain relationship in the overlying firn-pack has been presented elsewhere (Miller, 1956, ch. IX, c-5).

([8]) i. e. 10^6 dynes $cm^{-2} \simeq 1$ kg $cm^{-2} \simeq 1$ atmosphere.

([9]) Sometimes referred to as the «effective strain rate» (v. Nye, 1953, pp. 478-9).

forces (Nye, 1952) which have not been taken into account. It is not to be expected that line I is analogous since, as one proceeds down-glacier from site 10B, there are an increasing number of tension crevasses across the line of flow. Additionally, there is an increase of 0.4° in gradient at the 1953 bore site compared to 1950. These facts, and the apparent twisting of the pipe as previously discussed, imply that the bore-hole was moving from an area of some extension to greater extension. In the Jungfraufirn, the rate of extension was measured as 14 per cent per year. The close agreement between lines I*b* and II suggests that a similar rate of extension prevails in this sector of the Taku, where no such measurement has as yet been made. From these considerations it would appear that the pure shear analysis rather *over estimates* the relative importance of englacial shear stress. A comparison between theoretical and observed values is therefore useful for giving an estimate of the effects of extending and compressive flow.

For the basic comparison, n and k are derived from line IV and from lines I*a*, I*b* and I*c* in Figure 7. The value of n represents the slope of each line. The constant k is obtained from the fundamental stress/strain formula at the stress of 1 bar. Since the logarithm of 1 is zero, k is the ordinate at the point where the abscissa of any log stress/log strain rate line is zero. The constants in each of these cases and in the relationship given by line V are:

	n	k
Line I*a* (10B, 1950-51)	4	56.2
Line I*b* (10B, 1951-52)	1.6	0.4
Line I*c* (10B, *mean*, 1950-52	3	7
Line IV (Quasi-viscous creep rate, from laboratory tests)	4	0.25
Line V (Flow law, derived from tunnel measurements)	3	0.33

Although the mean value of n at site 10B is somewhat lower than the experimental quasi-viscous rate in compressive flow, it is in agreement with the exponent of effective shear stress found by Nye to pertain in the tunnel tests. This supports that the quantity $n = 3$ is of an order close to the average field condition usually found in low gradient temperate glaciers. Therefore, by assuming this value as reasonable, and by extrapolating the relationship of line I*c* all the way to the base of the glacier, a determination is made of the total flow *within the glacier*. (It is reiterated that this is on the ad hoc assumption that the simple power law for general analyses may be considered to govern the englacial deformation rate at depth.) The same is done for each set of data. For this calculation the difference is found between the *plastic flow velocity at the measured seismic depth* (U_D) and the *surface velocity* (U_o) at the top of the bore-hole. The relation is expressed by combining the empirical constants with the stress at this depth(D) in the manner suggested by Nye (1952, p. 86):

$$U_o - U_D = \frac{k}{n+1} D \tau^n$$

The Proportion of Basal Sliding

As far as the total transfer of ice across this profile is concerned, the differential flow within the glacier is seen to be small between the base of the pipe and bedrock. This is well demonstrated in part II of Table II, and is a fact directly related to the low gradient in this sector. The effect of increased slope is illustrated by the figures based on the quasi-viscous creep law, showing an increase of 1 to 36 per cent of englacial movement with a change of surface gradient of only 1° to 3°. Thus, in

theory, there is verification of the small mass transfer accredited to differential movement on the pipe profile — i.e. in the upper quarter of the glacier. In contrast, there is the substantial surface velocity which must be explained. Since the calculations in Table II are based on what is considered a reliable determination of depth, the *dominant translation of material is concluded to be from «sliding» of the glacier on its bed.* ([10]) On the Aletsch Glacier, near the Jungfrau bore-hole, the average gradient was 4.6°, with basal slippage calculated to account for about one-half of the surface velocity. On the Taku Glacier, which is similar in form, but with an average gradient of considerably less, a greater proportion of mass movement is by sliding. According to the mean calculated from the measured 1950-52 deformation (v. Part I of the table),

TABLE II

The Amount of Movement Within the Glacier According to The Measured and Theoretical Rates of Shear Strain

	Movement Within the Glacier		Shear Stress on Bed in bars	Basal Sliding, as Proportion of Total Movement
	(in meters)	(in % of total)		
I. According to extrapolated creep rate from pipe data				
1950-51 (line I*a*)	166	76	0.51	16%
1951-52 (line I*b*)	22	5	0.51	92%
Est. mean of I*a* I*b*	—	—	—	54%
Calculated mean over 1950-52 (line I*c*)	76.5	23	0.51	67%
II. According to quasi-viscous rate from laboratory tests				
assuming 1° slope, as on seismic profile	1	1—	0.51	99%
assuming 2° slope	17	7	1.0	93%
assuming 2.5° slope	34	15	1.2	85%
assuming 3° slope	84	36	1.5	67%
III. According to rate found in tunnel measurements				
assuming 1° slope, as on seismic profile	4	1	0.51	98%
assuming 2.5° slope	47	20	1.2	80%

([10]) The term «sliding», as hereafter used, is applied loosely since in some glaciers the movement can in large measure be due to intensified differential shearing in a thin zone of tectonic foliation close to the bed. The writer has observed such in a section one meter thick at the rock contact in the tunnel of Skautbreen, Norway (v. McCall, 1952, Fig. 5), and in sections up to 40 feet thick on certain Greenland glaciers. Such cases may be accentuated by a higher value of the k constant in the creep law, as a result of impurities in the water of the ice at the sole of the glacier (v. footnote page 446). The effect is the same, however, as if yielding were by pure slip at the bedrock interface.

two-thirds of the down-glacier movement at Station 10B is due to bottom slip, or to basal shearing. This leaves one-third consequent on flowage within the main body of the glacier. These ratios are, of course, based on the mean surface slope of 1°. At the same gradient, the theoretical values are compared, including those given by Nye's flow law (v. Parts II and III of Table II). On this hypothetical basis, 98 to 99 per cent of the mass movement would be due to bottom slip and even by using the average overall gradient of the main branch of the Taku Glacier (2.5°), 80 to 95 per cent would still be accorded to basal slip. On empirical grounds, such high ratios cannot be justified on Movement Profile IV (Fig. 3). Therefore, *the mean percentage of basal sliding calculated from the field measurements is considered to pertain. It is of interest that this value is 32 per cent less than that determined from the quasi-viscous rate.*

A Trajectory of Longitudinal Stress

The actual extent to which bottom sliding takes place depends not only on the surface gradient but on the slope and roughness of the underlying topography, on the width and thickness of the ice in question, and on any blocking factors involved, such as a tributary glacier abutting against a trunk glacier at the junction of two valleys. All these elements, of course, vary in different sectors of the icefield. Thus the amount of slippage on the glacier floor at the bore-hole is not to be construed as representative, even though for purposes of analysis this profile has been our essential concern. To illustrate this point a comparison is made with the transects on the main lower glacier. On Profile II an even higher ratio of sliding is calculated from the empirical law — i.e. 92 per cent. On another profile, two miles above the terminus, 54 per cent is indicated. The difference in these ratios may be explained by the narrower channel and increased depth of ice near sea-level.

Substantial slippage and marginal shearing also take place along the edges of the glacier. This is suggested by the form of the transverse surface velocity curves in Figure 3. Additionally in support of the effectiveness of true bottom slippage are the following facts: (1) the deeply entrenched and U-shaped channel of the lower glacier (Fig. 1, especially the sector down-glacier from Site 11*a*); (2) a noteworthy display of deep grooving and lateral striations along the bedrock margins; and (3) the vast quantity of rock flour which is constantly silting into Taku Inlet along the lines of the glacier's advancing front and which produces a characteristic milky color in the water of the fiord for some miles out. *None of these conditions could develop without considerable abrasion from the glacier's actually sliding on its bed.*

As for the basic cause of the high ratios of slippage described, it is clear that *flow within the glacier at a slope of 1° is so small that other stresses than those explained by pure shear must be in effect. Any important influence by hydrostatic pressure is ruled out, on the basis that it is as negligible in ice as in liquids.* ([11]) *The most significant supplemental stress can probably be attributed to a strong longitudinal force superimposed on the normal laminar flow and produced by an excessive increase in accumulation on the higher névés of the Taku-Llewellyn Glacier system in the past 100 years. To this stress tensor, and probably to an attendent increase in temperature within the ice itself, the phenomenal 20th-century advance of the Taku Glacier has been ascribed* (Miller, 1954*b*).

([11]) A conclusion cited as probable by Nye (1952, p. 82) and recently verified by Steineman (1956, personal communication) in laboratory tests at the Weissfluhjoch Research Station. Steinemann's experiments were made in a torsional-shear apparatus at a temperature of —1.9° C. and have shown that there is no influence whatsoever from *confining pressure* (up to 90 atmospheres) either on the actual flow rate in ice or on the accelerating or decelerating phases of recrystallization.

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THE MAIN PROBLEMS OF MODERN GLACIOLOGY IN THE LIGHT OF INVESTIGATIONS BY SOVIET SCIENTISTS

P. A. SHUMSKY

1. Formerly, glaciology was artificially confined to a study of glaciers. Modern glaciology studies all the ice on the globe — the Earth's cryosphere, comprising atmospheric, river, lake and sea ice, snow cover, glaciers and motionless ground, and underground ice.

2. Climate and ice formation cannot be treated as a cause and its consequence. Ice formation in all its varieties and climate are a result of heat and moisture exchange in the system: atmosphere—hydrosphere—lithosphere.

3. The nature of ice formation on the Earth's surface is, mainly, due to heating values and annual variations of heat loss, heat transfer and atmospheric precipitations, and in hydrosphere also by intensity of the water movement.

4. Among the processes of ice formation and transformation the following stages should be distinguished: origin of ice crystals, their growth, accumulation of ice crystals which are disconnected in the course of their growth, their transformation into a monolythic ice mass and subsequent changes of the latter.

5. There are three types of formation of monolythic ice masses:

a congelation — by water freezing;

b) recrystallization — by accumulation of ice crystals and their freezing together without thawing;

c) infiltration — by partial thawing, infiltration and freezing of melted water in pores of heaps of atmospheric ice crystals.

6. In conformity with regular vertical and latitudinal measurements of a heat and water balance of the Earth's surface, ice formation is distributed on the globe by zones. The Earth's cryosphere (except atmospheric ice) consists of two zones: the upper recrystallization and the lower congelation separated by an intermediate infiltration zone. Below the permanent part of cryosphere there is a seasonal congelation zone.

7. In regions of sea climate the congelation zone is narrow but in regions of sharply continental climate it replaces a probably disappearing recrystallization zone. Thus, cryosphere is of disymmetrical structure.

8. The main type of ice formation in hydrosphere is periodically cracking or disappearing ice cover, chiefly of congelation origin. The recrystallization zone is within the limits of the continent but for a few exceptions. Continental products of ice formation in the recrystallization and infiltration zones are glaciers, and in the congelation zone — motionless ground and underground ice.

9. Ice accumulating on the Earth's surface in the recrystallization and infiltration zones is passing into lower zones of cryosphere or into the sea owing to its plasticity.

10. In its movement glacial ice experiences various kinds of dynamic metamorphoses, which often brings about complex tectonic structures but which does not completely eliminate the initial structural features. The structure of the upper layer of glaciers and glacial covers in respect to the supply depends on the nature of the primary processes of ice formation and varies with different glaciological zones.

11. The temperature regimen of glaciers is extremely various and complex. It is determined by a surface regimen, different radiation and heat properties of snow cover, firn and ice, time variations of quite different mechanisms of heat loss in the

upper layers, a heat run-off, internal emission or tranfer of heat during its movement and the influence of the Earth's heat flux.

12. The character and intensity of geologo-geomorphological erosion and accumulation activity of glaciers depend on their morphological, dynamic and thermophysical properties.

13. Ice formation in the congelation zone exclusive of regions of the ablation of glaciers and glacial covers is represented by motionless ground ice of «snow areas» and hydroeffusions, and especially — by underground ice. The latter can be divided into three genetic groups:

1) congelation ice, formed by freezing underground waters contained in not yet frozen rocks;

2) cavern-lode ice formed from water and water vapours which penetrate into voids of frozen rocks;

3) buried ice, the product of buried ground ice.

14 The congelation ice comprises: concrete, lode, cegregation and injection ice, the cavern-lode.

Buried ice is of glacial origin only and, as a rule, does not exist within the boundaries of modern glacial regions.

15. In motionless ice of the congelation zone the substance is temporarily excluded from the general water circulation whereas in underground waters it can be preserved for a long geological period.

16. Underground ice and traces of its thawing is an important means in representing the ice formation history as traces of geological activity of glaciers and ice covers.

17. Cryosphere as a whole and its different parts constantly changing under the influence of general and local fluctuations of heat and moisture exchange. General changes are of main importance. They are likely to be caused by variations of solar activity through the intensification of the general circulation in atmosphere and hydrosphere.

18. Due to more considerable changes of the atmospheric circulation in maritime regions and the influence of the surface ice formation in the recrystallization and infiltration zones on a heat and moisture exchange, general changes of solar activity cause much greater movements of cryosphere in maritime regions than on the continent.

19. The invasion and retreat of ice formation take place simultaneously in both the hemispheres of the Earth. In the Antarctic cryosphere is at the lowest level, and the ice formation reaches the greatest size due to an optimum combination of oceanic conditions and the size of the continental foundation where the ice accumulates.

FEINBEWEGUNGSMESSUNGEN AN EINIGEN OSTALPEN-GLETSCHERN

GERHARD LINDIG
Frankfurt am Main

Abstract

During the years 1953 to 1956 precise measurements of the velocity by means of a theodolite were performed on three glaciers of the Œtztal Alps. The observations made at regular intervals of half an hour to one hour during day and night gave up to now the following results. The Gepatsch glacier shows a strictly linear movement at the point of observation whilst the Hintereis and the Gurgler glacier had in addition superposed pulsations. Retrograde and short, jerky motions have been observed.

1. Einleitung

Die Lösung der Fragen, ob die Feinbewegung des Eises:
a) kontinuierlich linear,
b) kontinuierlich periodisch,
c) ruckweise einsinnig in der Hauptfließrichtung oder
d) turbulent
ist, kann zur Klärung der allgemeinen Problematik der Gletscherdynamik, ob Block-, Viskositäts- oder Plastizitätstheorie gilt, beitragen.

In der Literatur finden sich Angaben über Messungen, die am Ende des vergangenen Jahrhunderts von Koch, Klocke und Pfaff durchgeführt worden waren [1], [2], [3] und [4]. Sie ermittelten unregelmäßig hin und her bezw. auf und ab führende Eisbewegungen. In neuerer Zeit beobachtete Bauer in Grönland [5] und Millecamps [6] am Mer de Glace turbulenzartige Gletscherbewegungen.

Die Messungsanordnung war mit Rücksicht auf den beschwerlichen Gerätetransport so einfach wie möglich vorgesehen, Mit einem Theodolit sollte am einen Gletscherufer in regelmäßigen Zeitabständen der Winkel gemessen werden zwischen einer oder mehreren Marken auf dem Eis (Eismarken) und einer festen Mire im gegenüberliegenden Fels.

a) Als Theodolit mußte ein möglichst genaues Instrument verwendet werden, welches jedoch nicht zu schwer sein durfte. Die Wahl fiel auf den Wild T2.

b) Der Instrumentenstandpunkt mußte nahe am Gletscherrand auf gewachsenem Fels gewählt werden und sollte möglichst tief liegen, um durch flache Visuren den Einfluß des Aufstellungsfehlers gering zu halten.

c) Das schwierigste Problem war bei der Aufstellung der Eismarken zu lösen da es fast unmöglich ist, einen Gegenstand über längere Zeit so fest mit dem Eis zu verbinden, daß er absolut ohne Eigenbewegung bleibt.

d) Die Mire als festes Anschlußziel mußte genau wie der Theodolit in gewachsenem Fels angebracht werden. Für die Nachtmessungen war, wie bei den Eismarken, Beleuchtungsmöglichkeit vorzusehen.

Sowohl günstige als auch ungünstige Verhältnisse wurden vorgefunden, wobei sich die überraschend geringe Gesamtbewegung der beobachteten Gletscher als besonders erschwerender Faktor herausstellte. Trotz gewisser Bedenken erklärte sich der Verfasser bereit, im Hinblick auf das Geophysikalische Jahr über die bisherigen noch nicht abgeschlossenen Untersuchungen zu berichten.

2. Gepatsch — Ferner 1953 [7], [8]

Die Station befand sich etwa 2 km oberhalb des Zungenendes, d.h. etwa 500 m unterhalb des großen Gletscherbruchs, auf der rechten Seite im « Prallhang » der gekrümmten Zunge. Die Situation war besonders günstig, da es möglich war, den Theodolit (Wild T2) in gleicher Höhe wie die Eismarke aufzustellen. Dadurch wurde erreicht, daß einesteils die Visierlinie nahezu horizontal verlief und andererseits Eismarke und Mire so eingerichtet werden konnten, daß sie gleichzeitig im Fernrohrgesichtsfeld erschienen. Die Winkelmessung konnte so mit dem Feintrieb allein vorgenommen werden. Bei allen Messungen wurden die Stativbeine *nicht* ausgezogen, um dem Theodolit weitgehendste Standfestigkeit zu verleihen und im Sitzen beobachten zu können.

Für die Aufstellung der Eismarke sollte ein neuer Gedanke erprobt werden. Die Ausrüstung bestand aus einem mit weißer Oelfarbe gestrichenen Holzbrett von $30 \times 30 \times 3$ cm³ Größe, in dessen Mitte die Marke aufgeschraubt war. Diese hatte die Form einer vertikal stehenden Scheibe von 10 cm Durchmesser mit je zwei diametral-liegenden weißen und schwarzen Quadranten, die nach Bedarf gedreht werden konnten. Im Mittelpunkt der Scheibe befand sich ein Loch von 1 cm Durchmesser unter Plexiglas, hinter welchem das Birnchen für die Nachtbeleuchtung angebracht war.

Das Brett wurde in eine genau nach seinen Maßen quadratisch ausgehauene Grube etwa 10 cm tief im Eis versenkt und wieder mit Eisstückchen zugedeckt, sodaß nur die Marke selbst herausragte. Mit Hilfe einer Dosenlibelle wurde das Brett möglichst genau horizontal eingerichtet. Der Anordnung lag der Gedanke zugrunde, daß das Brett nur vertikalen Druck auf das Eis ausübte und daher auch nur in dieser Richtung einschmolz, während seitliche Bewegungen nicht auftreten konnten.

Die Mire, welche als festes Anschlußziel in der gegenüberliegenden Felswand angebracht wurde, bestand aus der gleichen Scheibe mit Beleuchtungseinrichtung, die an einem kurzen Stab befestigt war und im Felsen mit Steinen verkeilt werden konnte. Drei gewöhnliche, flache Taschenlampenbatterien, parallel verbunden, lieferten eine Nacht hindurch genügend Strom für ein Lämpchen.

Die Messungen fanden vom 13. zum 14. 9. 1953 mit 4 Beobachtern in halbstündigem Rhythmus statt. Folgende Meßfolge war gewählt worden:

Mire — Marke; Mire — Marke; Mire — Marke.

Der Winkel γ zwischen Mire und Marke wurde aus den Mitteln der drei Ablesungen gebildet und außerdem zur Kontrolle des Bewegungsablaufes die Differenz

$$\Delta\gamma = \gamma_i - \gamma_{i-1}$$

Diese ist gleichzeitig proportional der jeweiligen Fließgeschwindigkeit im Halbstunden-Intervall. Sie war

$$\text{im Durchschnitt} \quad 25{,}3 \pm 6{,}0\ ^{cc}/\frac{h}{2}$$

$$\text{entsprechend} \quad 5{,}6 \pm 1{,}3^{mm}/\frac{h}{2}$$

mit den Extremwerten 14^{cc} und 42^{cc} pro halbe Stunde.

Der lineare Maßstab wurde durch Messen einer kurzen Vergleichsstrecke e bei der Marke und des dazugehörigen Winkels α in 5 Sätzen bestimmt.

Es ergab sich : $e = 64{,}8$ cm

$$\alpha = 2945^{cc} \pm 1{,}6^{cc}$$

und damit der Umrechnungsfaktor

$$k = \frac{e}{\alpha} = 0{,}22^{mm}/\text{cc}$$

sowie die Entfernung zur Marke

$$s = 140 \text{ m}$$

In der folgenden Tabelle 1 sind die beobachteten γ mit ihren zugehörigen Uhrzeiten und mittleren Fehlern, welche aus den Beobachtungsdifferenzen errechnet wurden, angegeben,

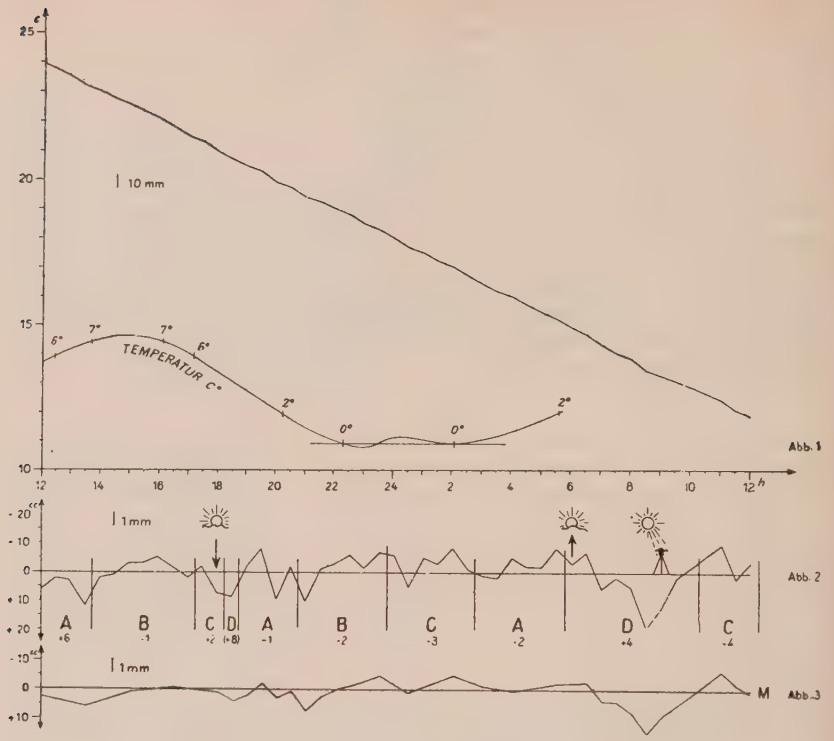
In Anlage I, Abb. 1 sind diese Werte graphisch aufgetragen. Sie zeigen einen überraschend guten, linearen Verlauf, wie er bisher noch nicht beobachtet wurde. Es läßt sich weder Abhängigkeit von der Lufttemperatur (am Instrument gemessen) noch von der Tageszeit feststellen, so daß die Theorie von den Tagesperioden zumindest für den speziellen Beobachtungsort als widerlegt gelten kann.

TABELLE 1

Beobachtungen Gepatsch-Ferner am 13. und 14.9.1953

Uhrzeit von	Uhrzeit bis	γ	Uhrzeit von	Uhrzeit bis	γ	Uhrzeit von	Uhrzeit bis	γ
h	h	g cc	h	h	g cc	h	h	g cc
12.00	12.07	0,2392 ± 2,5	20.30	20.43	0,1970 ± 3,3	5.00	5.13	0,1539 ± 2,7
12.30	12.40	0.2371 ± 8,3	21.02	21.11	0,1933 ± 2,7	5.30	5.39	0,1520 ± 5,3
13.00	13.05	0,2344 ± 2,9	21.30	21.38	0,1918 ± 3,8	6.00	6.10	0,1490 ± 4,8
13.32	13.38	0,2310 ± 5,5	22,00	22.06	0.1895 ± 2,9	6.34	6.42	0,1469 ± 3,9
14.00	14.10	0,2295 ± 5,3	22.30	22.42	0,1873 ± 3,7	7.03	7.08	0,1431 ± 1,8
14.30	14.37	0,2270 ± 5,0	23,00	23.08	0,1843 ± 1,4	7.31	7.38	0,1409 ± 2,6
15.00	15.06	0,2249 ± 4,2	23.30	23.35	0,1823 ± 2,1	8.00	8.10	0.1381 ± 2,4
15.30	15.38	0,2224 ± 4,1	24.00	24.12	0,1797 ± 6,3	8.30	8.40	0,1343 ± 3,7
16.00	16.09	0,2201 ± 5,7	0.30	0.43	0,1760 ± 4,4	9.01	9.10	0,1323 ± 3,3
16.30	16.37	0,2173 ± 1,3	1.00	1.10	0,1744 ± 4,2	9.30	9.40	0,1308 ± 5,0
17.00	17.07	0,2143 ± 2,4	1.30	1.39	0,1718 ± 4,1	10.00	10.07	0,1287 ± 3,9
17.30	17.38	0,2123 ± 3.1	2.00	2.09	0,1698 ± 3,0	10.30	10.37	0,1265 ± 11,0
18.00	18.12	0,2087 ± 5,6	2.30	2.40	0,1665 ± 3,7	11.00	11.05	0,1244 ± 4,1
18.30	18.38	0,2061 ± 3,9	3.00	3.20	0,1638 ± 2,3	11.30	11.34	0,1207 ± 3,5
19.00	19.10	0,2046 ± 4,7	3.30	3.42	0,1612 ± 2,2	12.00	12.04	0,1187 ± 5,6
19.31	19.41	0,2026 ± 4,6	4.00	4.12	0,1593 ± 4,9			
20.00	20.21	0,1984 ± 5,5	4.30	4.41	0,1564 ± 3,0			

Um zu klären, ob die überlagerten Unregelmäßigkeiten auf zufälligen Beobachtungsfehlern beruhten, oder ob sich Schwankungen in der Mikrobewegung nachweisen lassen, wurden in Anlage I, Abb. 2 die Abweichungen von der ausgleichenden Geraden in 10fachen Maßstab aufgetragen. Die Buchstaben A bis D geben die verschiedenen Beobachter an, darunter stehend die durchschnittlichen Abweichungen in den einzelnen Abschnitten. Eine merkbare Systematik in Abhängigkeit vom Beobachter ließ sich nicht feststellen.



Anlage I — Gepatsch-Ferner 1953

Die durchschnittliche Abweichung von der Geraden betrug $d\gamma = 4{,}9^{cc}$. Demgegenüber stand der Durchschnitt aller $m\gamma$ aus Tabelle 1 mit $\overline{m}\gamma = \pm\ 4{,}0^{cc}$, der sich zusammensetzte aus den Richtungsfehlern

für die Mire: $\overline{m}$ Mire $= \pm\ 2{,}7^{cc}$
für die Marke $\overline{m}$ Marke $= \pm\ 3{,}0^{cc}$

Der Unterschied zwischen $d\gamma$ und $\overline{m}\gamma$ *kann* in unregelmäßigen, vielleicht ruckweisen Eisbewegungen begründet liegen, es können aber auch Eigenbewegungen der Marke oder andere Fehlerursachen sein. Nur mit einer zweiten benachbarten Marke könnte dies geklärt werden.

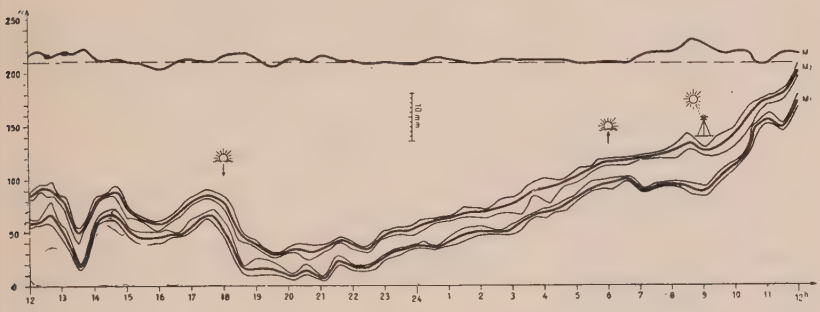
In Anl. I, Abb 3 ist der Versuch dargestellt, die Restkurve (Abb. 2) unter Berücksichtigung der mittleren Fehler der Einzelpunkte (Tab. 1) so weit wie diese es zulassen, zu linearisieren. Wenn dies Verfahren eine gewisse Zwangslösung darstellt, so zeigt es doch deutlich die Abweichungen, mit welchen *mindestens* gerechnet werden muß. Man sieht, daß sie sich auf eine Ausnahme (08.30 Uhr) in der Größenordnung eines Millimeters halten. Hier, wie bei allen späteren Untersuchungen wurde der Grundsatz verfolgt, daß, bis nicht das Gegenteil *zwingend* erwiesen war, lineare Bewegungen angenommen werden mußten; denn: Jeder lineare Verlauf wird durch zufällige Fehler in einen unregelmäßigen verfälscht, während das umgekehrte unwahrscheinlicher Zufall wäre.

im häufigeren Auftreten des Vorzeichens « Plus » in der Differenz:

3. Ablesung minus 1. Ablesung

für die Marke machte sich der Einfluß der Eisbewegung während der Messung bemerkbar.

In Anlage II wurden für die Mire M_1 und die Marke M_2 sämtliche Einzelbeobachtungen zu ihrem ungefähren Zeitpunkt aufgetragen (für letztgenannte die Werte



Anlage II — Gepatsch-Ferner 1953.

um den linearen Bewegungsteil reduziert) und die Einhüllenden um beide Punktscharen gezogen. Es zeigte sich ein merkwürdiges Verhalten der Beobachtungen zur Mire. Als Ursache dürfte in gewissem Umfang die Horizontalrefraktion in Frage kommen. Die Kurve M gibt den Abstand zwischen den beiden vermittelnden Linien M_1 und M_2 an und entspricht ganz der Abb. 3 in Anlage I (allerdings invers).

Ein Versuch zeigt, daß man *überall* zwischen die extremen Einhüllenden von M_1 und M_2 *konstanten* Abstand von 21^{cc} legen kann, mit Ausnahme des Abschnittes von 7.00 bis 9.00 Uhr. *Dort aber zeigt die Marke und damit der Gletscher weiter einen gleichmässigen Verlauf, während die Unregelmässigkeiten in den Visuren zur Mire liegen.* Refraktion, schlechte Zielauffassung oder ungünstige Zielbeleuchtung sind die möglichen Ursachen. Damit ist der Nachweis erbracht, daß die Messungen am Gepatsch — Ferner die *Annahme eines streng linearen Bewegungsablaufes* zulassen.

3. Hintereis — Ferner 1954

a) *Unternehmen Schimpp — Rudolph (17.-20.7. und 3.-5. 8.53)*

Die Station befand sich in der Felswand etwa 30 m über dem Gletscher auf dessen linker Seite, wenig unterhalb der Einmündung des Langtaufererl-Joch- Ferners. Durch diesen überhöhten Standpunkt wurden zwar Visuren dicht über dem Eis vermieden, aber es war nicht mehr möglich, Marke und Mire gleichzeitig ins Gesichtsfeld des Fernrohrs zu bringen. Die beiden Ziele wurden jedoch so eingefluchtet, daß die Kippung des Fernrohrs genügte, um von einem auf das andere Ziel überzugehen.

Als Eismarke wurde ein Vierkant-Stab von 1 m Länge verwendet, der zunächst straff in einem Bohrloch saß, im Laufe der Zeit jedoch etwas locker schmolz. Für die Nachtmessungen war an der, dem Instrument zugewandten Flachseite des Stabes eine Lampenfassung befestigt, die auch bei Tag angezielt werden konnte. Eine ebensolche Einrichtung war auch als Mire im gegenüberliegenden Fels angebracht worden. Die Entfernung zwischen Instrument und Eismarke betrug

$$e = 376 \text{ m und damit } k = 0{,}59^{mm}/cc$$

Da die drei Beobachter erst Erfahrungen sammeln mußten, war keine einheitliche Meßfolge angewendet worden. Sowohl die Zeitintervalle und Satzzahlen als auch die Beobachterfolge wurden öfter gewechselt, wodurch sich die Auswertung erschwerte.

Die Beobachterwechsel wirkten sich auf den mittleren Winkelfehler aus, der im Durchschnitt

$$\bar{m}\gamma = \pm 5^{cc}$$

betrug. Es ließen sich deutliche systematische Unterschiede zwischen den einzelnen Beobachtern feststellen. Aus den Messungen, die B und C gemeinsam durchführten, ließ sich ermitteln:

Beobachter B: syst. Fehler $+ 7^{cc}$ ($n = 29$)
Beobachter C: syst. Fehler $- 9{,}5^{cc}$ ($n = 23$)

Dies zeigt, dass bei Beobachterwechseln besondere Vorsicht geboten ist und gegebenenfalls mit grösseren Abweichungen gerechnet werden muss.

Die Messungen vom 3. bis 5.8.54 wurden auf Grund der Erfahrungen aus dem ersten Unternehmen mit homogeneren Beobachtungsreihen durchgeführt. Dadurch war der mittlere Winkelfehler mit

$$\bar{m}\gamma = \pm 3{,}1^{cc}$$

wesentlich kleiner.

TABELLE 2

Beobachtungen Hintereis-Ferner vom 17.—20. 7. 1954

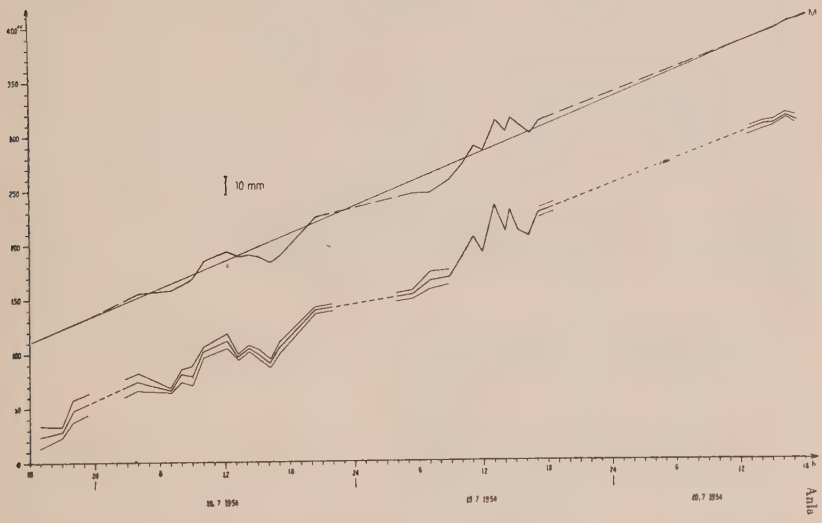
Uhrzeit Mittel	γ Mittel	Beobachter	n	Uhrzeit Mittel	γ Mittel	Beobachter	n
h	cc			h	cc		
18.45	264 ± 9.0	A	4	5.20	392 ± 4,0	B,C	6
21.00	268 ± 5,5	B,A	6	7.00	405 ± 8,0	B,C	7
22.0	286 ± 10,0	B,A	2	8.40	408 ± 7,0	C,B	5
1.00	265	B	1	9.40	424 ± 2,5	C,B	5
4.00	314 ± 8,0	A	2	11.00	446 ± 5,0	C,B	2
7.00	306 ± 2,0	A	2	11.40	431 ± 4,0	B	4
8.00	321 ± 6,0	A,B	3	13.00	475 ± 10,0	C	2
9.0	319 ± 9,0	A,B	4	14.00	449 ± 3,0	B	3
10.00	342 ± 4,0	A,B	4	14.20	470 ± 3,0	C	3
12.10	351 ± 7,0	B,A,C	4	15.10	451 ± 8,0	C,B	4
13.10	336 ± 3,0	B,A,C	5	16.10	446 ± 5,0	B,C	9
14.10	344 ± 3,0	A,C	3	17.00	467 ± 4,0	C	4
15.10	338 ± 6,0	C,B	3	14.00	547 ± 3,5	C	4
16.10	330 ± 4,0	B,C	4	15.00	549 ± 3,0	C	4
17.10	344 ± 5,0	C,B	4	16.15	556 ± 2,0	C	6
20.20	379 ± 3,0	C,B	3	17.15	552 ± 4,0	C	7

TABELLE 3

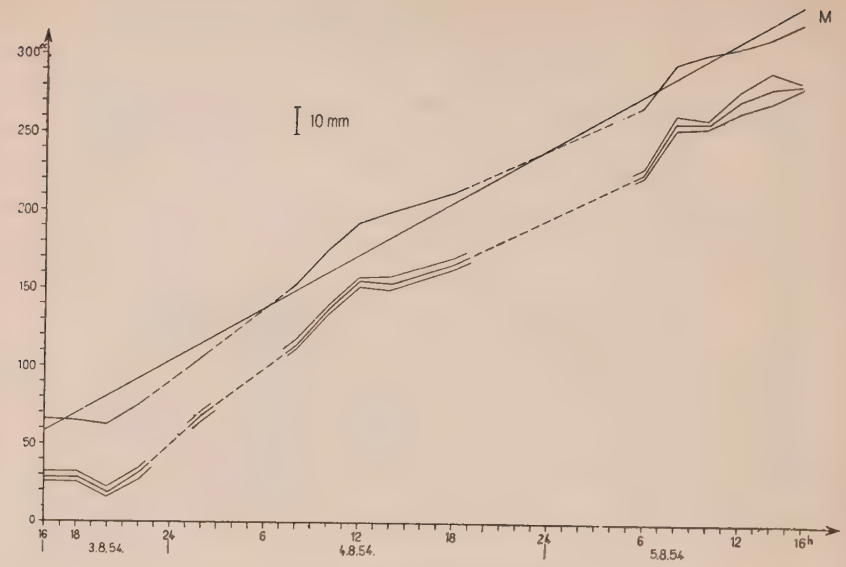
Beobachtungen Hintereis-Ferner vom 3.—5. 8. 1954

Uhrzeit Mittel	γ Mittel	Beob-achter	n	Uhrzeit Mittel	γ Mittel	Beob-achter	n
h	cc			h	cc		
16.00	729 ± 2,7	A	4	14.00	852 ± 5,1	A	4
18.00	729 ± 1,3	A	5	18.00	864 ± 2,9	A	5
20.00	719 ± 1,6	A	3	6.00	921 ± 1,1	A	4
22.00	731 ± 3,2	A	5	8.00	954 ± 4,3	A	5
2.00	767 ± 1,8	A	5	10.00	954 ± 2,0	A	4
8.00	813 ± 1,7	A	5	12.00	968 ± 5,9	A	5
10.00	836 ± 0,9	A	4	14.00	976 ± 11,5	A	2
12.00	853 ± 2,4	A	5	16.00	975 ± 1,2	A	4

Vorstehende Zahlenwerte sind in den Anlagen III und IV graphisch dargestellt.



Anlage III — Hintereis-Ferner (Schimpp) 1954.



Anlage IV — Hintereis-Ferner (Schimpp) 1954.

b) *Unternehmen Lindig*

Da die eben beschriebenen Untersuchungen im Gegensatz zu denjenigen am Gepatsch-Ferner wesentlich größere Abweichungen vom linearen Verlauf zeigten, deren Ursache in der Anlage der Messungen vermutet wurde, sollte am 5. and 6. 9.54 an der selben Stelle nochmals eine Beobachtungsreihe durchgeführt werden.

Es war vorgesehen, Messungen vom Stativ mit solchen vom Pfeiler zu vergleichen. Zu diesem Zweck wurde mit Zement aus Felsbrocken ein etwa 1 m hoher Beobachtungspfeiler aufgemauert. In beiden Fällen wurden Wild T2 — Theodolite benutzt, Sonnenschirme verwendet und wiederum im Sitzen beobachtet.

Um das Verhalten einer Stabmarke I im Gegensatz zur Brettmarke II zu klären, wurden beide im Abstand von etwa 2 m hintereinander aufgestellt. Die Zielausbildung beim Brett hatte in diesem Fall noch nicht die endgültige, bei den Gepatschmessungen beschriebene Form, sondern bestand nur aus einer Fassung, in die je nach Bedarf ein Birnchen oder ein kleiner Holzkegel von 2 cm Seitenlänge eingeschraubt werden konnte.

Die Messung erfolgte so, daß in einstündigem Rhythmus Ablesungen in drei Sätzen gemacht wurden, und zwar zuerst am Pfeiler und dann am Stativ, jeweils vom selben Beobachter. Die Meßfolge war

Felsmire — Stabmarke — Brettmarke
Brettmarke — Stabmarke — Felsmire
Felsmire — Stabmarke — Brettmarke

Vor Beginn jeder Messung wurden vorhandene Ausschläge der Alhidadenlibelle mit den Fußschrauben sorgfältig beseitigt, während sie nach der Messung zur späteren Berücksichtigung registriert wurden.

TABELLE 4

Beobachtungen Hintereis-Ferner am 5. und 6. 9. 1954

Uhrzeit	Pfeiler		Stativ		Beobachter
	Stab I	Brett II	Stab I	Brett II	
h	cc	cc	cc	cc	
13.00	720 ± 1,8	720 ± 1,3	720 ± 1,6	720 ± 1,8	A
14.00	730 ± 2,8	731 ± 2,5	723 ± 4,1	728 ± 3,4	A
15.00	733 ± 3,1	733 ± 2,9	733 ± 4,5	739 ± 4,8	A
16.00	735 ± 5,2	745 ± 4,7	736 ± 2,2	742 ± 3,9	A
17.00	741 ± 2,4	748 ± 2,4	739 ± 3,5	751 ± 6,1	A
18.00	742 ± 2,1	750 ± 3,3	743 ± 4,5	750 ± 3,6	B
19.00	(747)	(753)	zu dunkel		A
20.00	737 ± 0,9	750 ± 2,5			A
21.00	739 ± 2,5	758 ± 3,4			A
22.00	741 ± 0,9	771 ± 1,8			C
23.00	748 ± 2,4	775 ± 2,2			C
24.00	746 ± 4,5	763 ± 4,5			B
1.00	753 ± 1,9	762 ± 2,1			B
2.00	759 ± 2,0	770 ± 2,0			A
3.00	763 ± 1,7	776 ± 2,4			A
4.00	767 ± 2,3	782 ± 2,0			A
5.00	771 ± 2,7	784 ± 2,2			A
6.00	777 ± 4,1	779 ± 4,0			A
7.00	774 ± 3,0	784 ± 3,4	791 ± 2,7	795 ± 3,7	A
8.00	779 ± 3,8	777 ± 3,0	780 ± 10,8	791 ± 10,0	B
9.00	779 ± 3,8	776 ± 1,6			B
10.00	783 ± 2,3	794 ± 3,3			B
11.00	792 ± 3,2	799 ± 4,0	800 ± 4,7	810 ± 5,2	A

Im Durchschnitt ergaben sich folgende mittleren Fehler:

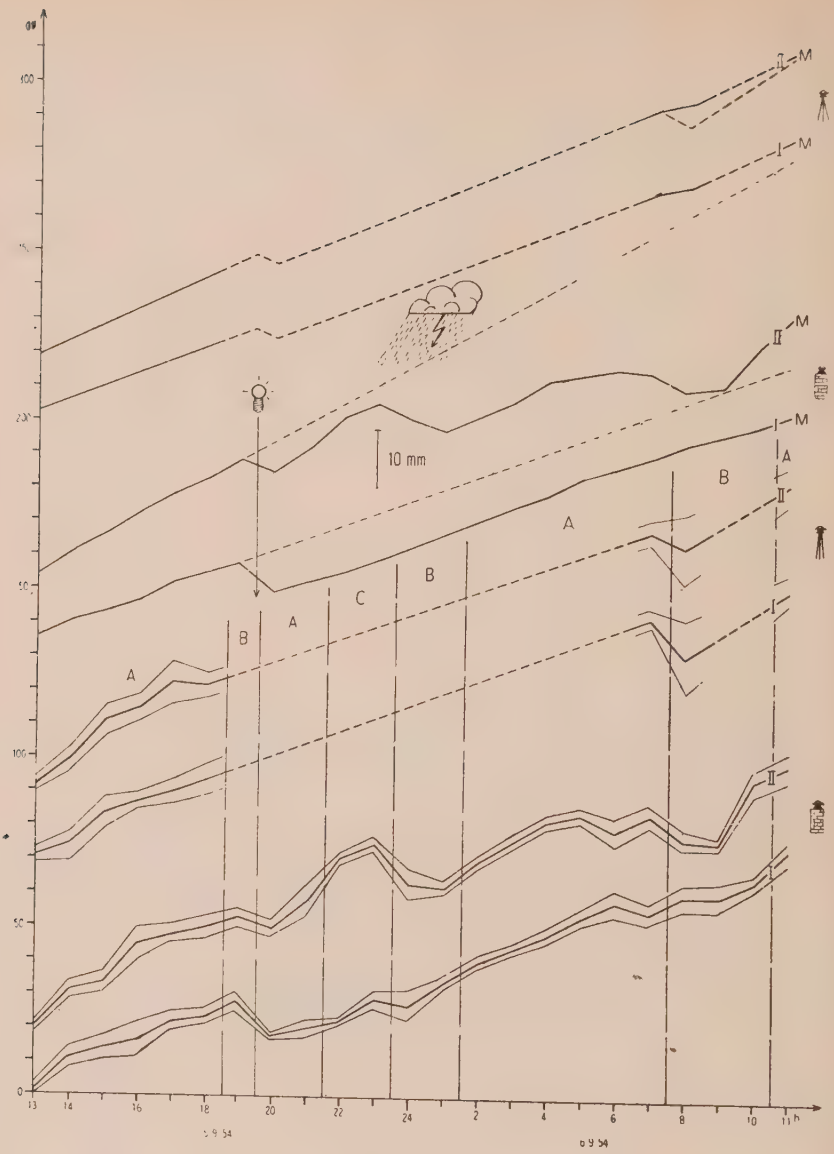
TABELLE 5

Durchschnittlicher, mittlerer Fehler

Stand	Mire	Stab	γ Stab	Brett	γ Brett	n
	cc	cc	cc	cc	cc	
Pfeiler	± 1,9	± 1,8	± 2,7	± 1,8	± 2,8	22
Stativ	± 3,0	± 2,4	± 4,3	± 3,6	± 4,7	9
Stativ red.	± 2,5	± 1,7	± 3,5	± 3,1	± 4,2	8

Die letzte Zeile « Stativ red. » zeigt Werte, die ohne die ausgesprochen großen Fehler beim Beobachterwechsel um 8.00 h berechnet wurden.

Die Ergebnisse der Tabelle 4 sind in der Anlage V graphisch dargestellt.



Anlage V — Hintereis-Ferner (Lindig) 1954.

Diskussion der Ergebnisse:

a) *Anlage III, Messungen vom 17.—20.7.54*

Im Gegensatz zu den Beobachtungen am Gepatsch-Ferner sind dem im ganzen linearen Bewegunsablauf Unregelmäßigkeiten überlagert, die um das 10fache größer

sind als dort. Es fragt sich, in wie weit die bereits geschilderte, inhomogene Beobachtungsanordnung, vor allem die ständigen Beobachterwechsel die Ursache dafür waren. Auffallend ist in dieser Hinsicht das letzte Stück der Kurve (20.7. 14-17 h)) welches nur von einem Beobachter stammt (vgl. Tabelle 2) und ruhiger verläuft. Um die selbe Zeit an den beiden Vortagen sind die stärksten Versprünge festzustellen, doch sind die Nachtbeobachtungen zu lückenhaft, um wirkliche Gegensätze verdeutlichen zu können.

Auch hier wurde wieder die mittlere Fehler-Einhüllende gezeichnet, die zur besseren Übersichtlichkeit für die Zeit 19.7 9-17 h weggelassen werden mußte. Hieraus konnte die mögliche Glättungskurve M entnommen werden, welche zeigt, daß nach Ausschaltung der reinen Beobachtungsfehler noch mit Unregelmäßigkeiten in der Größenordnung von 10 mm gerechnet werden muß. Rückläufige Bewegungen sind in vier Fällen als möglich anzunehmen. Da keine Vergleichsbeobachtungen vorliegen, ist nicht zu entscheiden, welcher Anteil eventuell der gelockerten Stabmarke zuzuschreiben ist.

b) *Anlage IV, Messungen vom 3.-5.8.54*

Erwartungsgemäß zeigt sich entsprechend der einheitlicheren Beobachtungsanordnung ein wesentlich ruhigerer Verlauf der Kurve. Obwohl wiederum erhebliche Nachtlücken zu verzeichnen sind, lassen die vorhandenen Beobachtungen linearen Verlauf in dieser Zeit vermuten. Bei Betrachtung der Glättungskurve M läßt sich keine rückläufige Bewegung als zwingend entnehmen. Auffallend sind die beiden Maxima etwa um die Mittagszeit herum, die mit den entsprechenden am 18. und 19.7. (Anlage III) zu korrespondieren scheinen. Andererseits kann man auch die ganze Kurve in zwei lineare Teilstücke mit Knickpunkt am 4.8 um 12 h zerlegen, die einer Geschwindigkeitsänderung von $V_1 = 112$ mm/Tag auf $V_2 = 68$ mm/Tag entsprechen. Inwieweit diese Erscheinung vielleicht einem allmählichen Neigen des Markenstabes zugeschrieben werden kann, oder ob es tatsächlich unterschiedliche Fließgeschwindigkeiten sind, läßt sich mit dem vorliegenden Material nicht entscheiden. Die Forderung nach mindestens zwei Marken, die sich gegenseigtig kontrollieren, wird hierdurch erneut bestätigt.

c) *Anlage V, Messungen vom 5.-6.9.54*

Die Zeichnung zeigt von ober nach unten :

Glättungskurven M	Stativ,	Brettmarke = II
	»	Stabmarke = I
	Pfeiler,	Brettmarke = II
	»	Stabmarke = I
Originalkurven mit Einhüllenden,	Stativ,	Brettmarke = II
	»	Stabmarke = I
	Pfeiler,	Brettmarke = II
	»	Stabmarke = I

Zunächst zeigen bis 19 h alle Beobachtungsreihen sehr befriedigende Übereinstimmungen. Allerdings fällt auf, daß die Geraden von Stab und Brett deutlich divergieren. Es besteht entweder die Möglichkeit, daß die Fließgeschwindigkeit auf so kurze Entfernung von 2 m zwischen $V_I = 50$ mm/Tag und $V_{II} = 77$ mm/Tag differiert hat, oder der Stab hat sich langsam aber stetig entgegen der Fließrichtung geneigt.

Zwischen 19 h und 20 h erfolgte das Auswechseln der Zielkegel gegen kleine Birnen, wobei beide Marken gegen die Vorwärtsbewegung bis zum Anschlag angedrückt wurden. Der Knick in den Kurven ist deutlich zu erkennen. Von 20 h an liefen zunächst wieder beide Marken linear divergent bis 23 h. Etwa ab 24 h setzte ein

heftiges einstündiges Gewitter mit Sturmböen, Regen und Hagelschauer ein, doch konnten die Beobachtungen dank der gut verankerten Schirme ohne Unterbrechung fortgeführt werden. Von diesem Zeitpunkt an pendelte die Brettmarke im Gegensatz zum Stab mit einer Amplitude von etwa 7 mm hin und her. Eine möglicher Erklärung ergibt sich aus folgender Feststellung beim Abbau der Station : Die Grube im Eis zur Aufnahme der Brettmarken war bis zum Rand mit Schmelz- (oder Regen-) wasser gefüllt, so daß das Brett darin hin und her schwimmen konnte. *Bei Verwendung einer Brettmarke muss daran gadacht werden, eine Abzugsrinne für das Schmelzwasser auszuschlagen.*

Der Genauigkeitsgewinn durch die Pfeilermessung wurde bereits in Tabelle 5 demonstriert, so daß die Feststellung getroffen werden kann : *Wenn es die zeitlichen und sonstigen Umstände erlauben, lohnt es sich wohl, 10 kg Zement zum Bau eines Pfeilers auf die Station zu transportieren.*

4. Gurgler Ferner 1955

Drei verschiedene Eismarken waren hergestellt worden, um die beste Lösung für deren Aufstellung ermittlen zu können. Folgende Kombinationen wurden verwendet :

I. Kegel : Diese Marke bestand aus dem (Seite 3) erwähnten 30 × 30 cm² Brett, auf welchem in der Mitte ein Holzkegel von 10 cm Seitenlänge als Ziel stand. Die Spitze dieses Kegels war aus Plexiglas gedreht und konnte abgenommen werden. Darunter befand sich ein Hohlraum in welchem das Birnchen für die Nachtbeleuchtung saß. Halb weißer und halb schwarzer Oelfarbenanstrich erlaubte die Auswahl des günstigsten Kontrastes.

II. Stab : Die zweite Marke bestand aus einem weißgestrichenen Hartholzstab von 1 m Länge und 3 cm Durchmesser, an dessen oberem Ende die bereits (Seite 3) beschriebene Zielscheibe befestigt war. Das Bohrloch wurde mit einem Eisbohrer, bestehend aus einem 1,50 m langen Stahlrohr mit 4-5 scharfen Zähnen am unteren Ende, in etwa 5 Minuten hergestellt.

Um den Stab herum wurde aus faustgroßen Steinen eine « Künstliche Moräne » aufgeschüttet um die Abschmelzung zu verringern.

III. Brett : Hiermit sei in Kurzform die Eismarke bezeichnet, wie sie schon (Seite 3) beschrieben wurde. Sie bestand aus einem Grundbrett mit vertikaler Zielscheibe.

Diese drei Marken waren im Abstand von etwa 1 m in Dreiecksform aufgestellt und gegen die Mire eingefluchtet worden.

Die Entfernung betrug $e = 543$ m von der Station, die sich unmittelbar von der Westwand des kleinen Zollhäuschens unweit des Hochwildehauses befand.

Die Messungen wurden am 2.-3.9.55 in einstündiger Folge mit zwei Theodoliten Wild T2 auf Stativ nacheinander, teilweise mit verschiedenen Beobachtern und z.T. durch denselben Beobachter, durchgeführt. Außerdem wurde ab 12.15 h an einem Instrument (Th.N.) mit Sonnenschirm und am anderen (Th.S) ohne einen solchen gemessen, um den Einfluß auf die Meßgenauigkeit festzustellen. Die Beobachtungszeit pro Durchgang betrug 5-7 Minuten. Es war die Frage, ob sich auch bei der sehr geringen Durchschnittsgeschwindigkeit von 18 mm pro Tag die Eisbewegung während der Messung bemerkbar machen würde. Hierzu gaben die Beobachtungen von 9 h am Theodolit Süd (Th. S.) aufschlußreichen Einblick. In der folgenden Tabelle 6 sind diese Einzelablesungen wiedergegeben. Die Meßfolge war :

1. Zeile : von rechts nach links
2. Zeile : von links nach rechts und
3. Zeile : von rechts nach links

TABELLE 6

Auszug aus dem Messungsprotokoll vom 3.9.1955

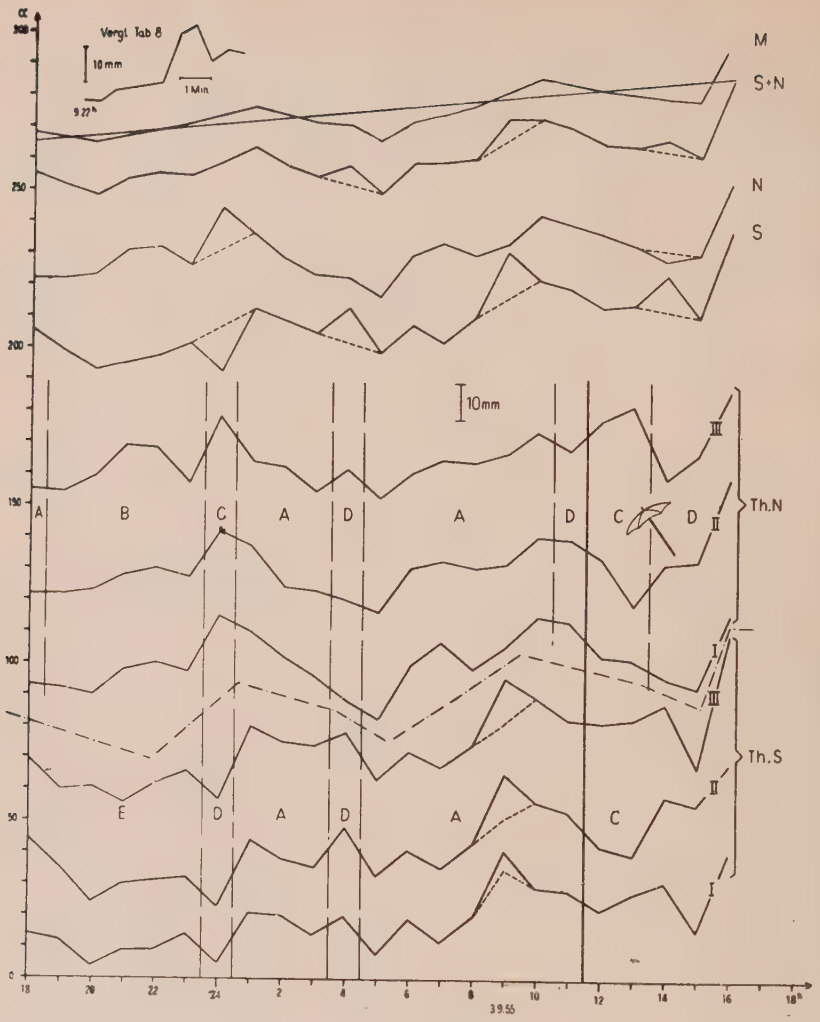
Instrument: Wild T2 (Stativ Süd) k = 0,85 mm/cc

Uhrzeit von	Uhrzeit bis	Mire g	Mire cc	Kegel g	Kegel cc	Stab g	Stab cc	Brett g	Brett cc	Beobachter	Bemerkungen
h	h										
9.22	9.28	1,1729	+ 2	1,2585	+ 7	1,2582	+14	1,2252	+14	A	Instrument in der Sonne
		728	+ 3	587	+ 5	602	— 6	274	— 8		
		736	— 5	604	—12	605	— 9	271	— 5		
Mittel:		1,1731 ± 2,5		1,2592 ± 6,0		1,2596 ± 7,2		1,2266 ± 6,8			
				861 ± 6,5		865 ± 7,6		535 ± 7,3			

TABELLE 7

Beobachtungen Gurgler-Ferner am 2. und 3.9.1955

Theodolit Nord				Theodolit Süd			
Uhrzeit h	I Kegel cc	II Stab cc	III Brett cc	Uhrzeit h	I Kegel cc	II Stab cc	III Brett cc
18.15	100 ± 2,1	100 ± 2,0	100 ± 1,7	18.20	100 ± 4,5	100 ± 4,7	100 ± 4,5
19.15	99 ± 2,9	100 ± 2,7	99 ± 1,8	19.20	98 ± 3,0	91 ± 4,5	91 ± 3,4
20.20	97 ± 3,7	101 ± 1,8	104 ± 2,1	20.30	90 ± 3,2	80 ± 2,6	92 ± 3,7
21.13	105 ± 5,0	106 ± 3,4	114 ± 3,2	21.20	95 ± 3,5	86 ± 3,4	87 ± 2,3
22.14	107 ± 3,2	108 ± 3,4	113 ± 2,0	22.21	95 ± 3,3	87 ± 3,3	93 ± 3,2
23.13	104 ± 1,8	105 ± 2,2	102 ± 1,3	23.20	100 ± 1,0	88 ± 1,5	97 ± 1,9
0.40	122 ± 5,0	120 ± 5,3	123 ± 5,9	0.20	91 ± 4,9	79 ± 3,6	88 ± 4,0
1.15	117 ± 5.6	115 ± 8,3	109 ± 7,6	1.30	107 ± 5,4	100 ± 3,8	111 ± 3,3
2.15	109 ± 2,3	102 ± 2,3	107 ± 4,0	2.23	106 ± 4,3	94 ± 4,3	106 ± 4,6
3.15	103 ± 5,5	101 ± 5,5	99 ± 6,0	3.25	100 ± 5,4	91 ± 4,1	105 ± 2,2
4.15	95 ± 4,1	98 ± 4,2	106 ± 3,7	4.22	106 ± 6,5	104 ± 6,1	109 ± 7,3
5.17	89 ± 4,8	94 ± 3,3	97 ± 4,8	5.24	94 ± 6,5	89 ± 3,2	94 ± 1,8
6.15	106 ± 6,0	108 ± 4,8	105 ± 3,7	6.34	105 ± 4,4	97 ± 3,5	103 ± 2,8
7.15	114 ± 1,4	110 ± 2,9	109 ± 2,2	7.21	98 ± 2,9	91 ± 2,9	98 ± 3,1
8.15	105 ± 1,0	108 ± 0,8	108 ± 1,9	8.23	106 ± 4,1	99 ± 2,8	105 ± 3,6
9.15	112 ± 2,7	109 ± 1,8	111 ± 1,8	9.22	127 ± 6,5	121 ± 7,6	126 ± 7,3
10.13	122 ± 1,8	118 ± 3,6	118 ± 2,6	10.21	115 ± 2,9	112 ± 4,6	120 ± 2,9
11.15	120 ± 2,6	117 ± 3,7	112 ± 6,7	11.22	114 ± 5,5	109 ± 8,7	113 ± 5 5
12.15	109 ± 4,8	111 ± 5,1	122 ± 5,2	12.25	108 ± 4,1	98 ± 4,5	112 ± 3,2
13.20	108 ± 4,0	96 ± 2,2	126 ± 4,5	13.25	113 ± 5,8	95 ± 5,0	113 ± 5,1
14.15	102 ± 5,2	109 ± 6,9	103 ± 5,1	14.40	117 ± 3,0	114 ± 1,9	118 ± 3,2
15.15	99 ± 5,7	110 ± 6,7	111 ± 5,0	15.25	101 ± 2,7	111 ± 5,1	98 ± 5,6
16.20	122 ± 5,5	136 ± 3,4	131 ± 3,3	16.25	126 ± 3,3	124 ± 5,5	140 ± 3,8



Anlage VI — Gurgler-Ferner 1955.

Wie man deutlich erkennt, muß der Gletscher in der kurzen Zeitspanne zwischen der zweiten Ablesung zum Kegel (587) und der zweiten zum Stab (602) *einen Ruck von etwa 20cc bezw. 17 mm* (die gesamte Tagesbewegung) ausgeführt haben. Die kleine Detailskizze oben links in Anlage VI zeigt den Bewegungsablauf, wie er sich abgespielt haben könnte, wenn nicht zufällig, außergewöhnliche *Refraktionsverhältnisse* diese merkwürdigen Beobachtungsergebnisse verursacht haben, da die dritte Ablesung der Mire einen, wenn auch kleinen Versprung von 8cc zeigt. Ähnliche, jedoch etwas geringere Veränderungen als in Tabelle 6 traten bereits um 1.15 Uhr nachts am Theodolit Nord auf. Als Konsequenz wurde die Meßfolge ab 12.15 Uhr geändert, indem jedes Ziel sofort nacheinander drei Mal eingestellt und abgelesen wurde, um die Zeitspanne pro Ziel möglichst kurz zu halten.

Die Durchschnittswerte der mittleren Fehler zeigt Tabelle 8

TABELLE 8

Stand	Mire	Kegel			Stab			Brett		
	m_r	m_r	m_γ	m_m	m_r	m_γ	m_m	m_r	m_γ	m_m
	cc	cc	cc	cc	cc	cc	cc	cc	cc	cc
Th. N	± 2,5	± 2,6	± 3,8	± 3,4	± 2,5	± 3,8	± 3,6	±2,5	± 3,8	± 6,0
Th. S	± 2,8	± 2,8	± 4,1	± 2,5	± 2,9	± 4,2	± 3,7	± 2,3	± 3,8	± 3,3

Hierin bedeutet :

m_r = Fehler der Richtung zu den einzelnen Zielen
m_γ = Fehler des Winkels zwischen Mire und Marke
m_m = Fehler der einzelnen Marke gegenüber dem Mittel aus drei Marken.

Anlage VI zeigt neben den Einzelbeobachtungen : dieses Mittel S aus den drei Markenablesungen am Theodolit Süd, dasselbe Mittel N für Theodolit Nord, die Kurve S + N als das Mittel aus S und N und schließlich die Glättungskurve M. Für die jeweiligen drei Messungen mit Theodolit S bezw. Theodolit N lassen sich gewisse Parallelitäten erkennen. Auffallend ist, daß Th. S. ein viel unruhigeres Bild lieferte als Th.N., was mit den größeren Fehlern in Tabelle 8 übereinstimmt. Wie die mittleren Abweichungen m_m zeigen, die in ihrer Größenordnung mit den m_γ übereinstimmen, meist sogar kleiner sind, kann angenommen werden; daß die Differenzen zwischen den drei Marken im wesentlichen auf zufälligen Meßfehlern beruhen. Vergleicht man allerdings in den Zeitpunkten 14^h und 15^h die Kurve Th. S II mit Th. N II, so ist dort eine Eigenbewegung der Marke oder der örtlichen Eispartie sehr wahrscheinlich.

Ein etwas klareres Bild liefert der Vergleich der beiden Mittelkurven S und N, die gewisse Parallelitäten erkennen lassen, obwohl noch eine ganze Reihe Diskordanzen übrig bleiben. Diese verringern sich allerdings beträchtlich, wenn man einige Außenseiter wegläßt, die in drei Fällen durch Beobachterwechsel und um 9 h durch den oben besprochenen « Ruck » bedingt sein können. Der durch die gestrichelten Kurvenstücke dargestellte Verlauf läßt bereits eine gemeinsame Tendenz erkennen.

Die restliche Diskrepanz kann im Nacheinander der Messungen an den beiden Instrumenten begründet liegen. Dies gewinnt dann sehr an Wahrscheinlichkeit, wenn man die Möglichkeit kurzfristiger, sprunghafter Bewegungsvorgänge anerkennt. Besonders bemerkenswert ist in diesem Zusammenhang der eindeutige Sprung von durchschnittlich $25^{cc} = 21$ mm, den alle 6 Beobachtungen zwischen 15 h und 16 h zeigen.

Über die Kurve der Gesamtmittel S + N entstand schließlich die Glättungskurve M, wobei ein mittlerer Fehler von 2^{cc} zugrunde gelegt wurde. Auch hier ist eine Art pulsierender Bewegung mit einer Amplitude von etwa 10 mm übrig geblieben. Dieselbe Größenordnung wie beim Hintereis-Ferner ! Dieses Pulsieren scheint demnach unabhängig von der Fließgeschwindigkeit zu sein.

TABELLE 9

Beobachtungen Gurgler-Ferner vom 22.—26.8.1956

Datum	Uhrzeit	Wild T3 (Pfeiler)				Wild T2 (Pfeiler)			
		Kegel	Stab	Brett	Beob.	Kegel	Stab	Brett	Beob
	h	cc	cc	cc		cc	cc	cc	
22.8.	18.30	100 ± 3,3	100 ± 3,2	100 ± 2,8	A				
	23.04	101 ± 3,0		100 ± 3,1	A				
23.8.	1.02	98 ± 2,0							
	14.12	138 ± 8,5	136 ± 3,0	136 ± 2,3	A				
	15.00	123 ± 1,6	137 ± 1,9	135 ± 1,8	A				
24.8.	7.40	140 ± 0,8	146 ± 1,6	150 ± 1,6	A				
	8.41	140 ± 0,8	148 ± 1,6	148 ± 1,6	A				
	9.42	141 ± 0,9	150 ± 1,2	150 ± 0,9	A				
	10.42	141 ± 1,5	154 ± 1,3	155 ± 2,1	A				
	11.41	145 ± 1,3	154 ± 1 3	164 ± 2.1	A				
	12.41	143 ± 1,5	157 ± 1,2	156 ± 1,5	A				
	13.42	147 ± 1,7	161 ± 1,8	156 ± 1,8	A				
	14.42	153 ± 0,7	163 ± 0,8	168 ± 1,2	A				
	15.48	144 ± 1,3	157 ± 1,4	152 ± 1,2	A				
	16.45	144 ± 1,2	154 ± 1,7	158 ± 1,2	A				
	17.44	144 ± 0,9	158 ± 1,2	162 ± 1,0	A				
	20.00	146 ± 1,3	156 ± 1,5	162 ± 1,3	A				
	21.00	145 ± 1,8	152 ± 3,2	157 ± 1,5	A				
	22.00	135 ± 1,9	145 ± 0,8	144 ± 0,8	A				
	23.00	145 ± 0,9		153 ± 0,9	A				
25.8.	0.00	142 ± 1,0		158 ± 1,0	A				
	1.00	147 ± 0,9		166 ± 0,9	A				
	2.00	153 ± 1,0		170 ± 1,4	A				
	3.00	157 ± 0,9		173 ± 0,9	A				
	4.00	150 ± 1,2		159 ± 0,9	A				
	5.00	148 ± 1,4	163 ± 1,5	157 ± 1,3	A				
	6.00	148 ± 0,6	175 ± 0,7	175 ± 0,7	A				
	7.00	154 ± 0,7	168 ± 0,8	171 ± 1,0	A				
	8.00	150 ± 1,3	166 ± 1,3	174 ± 1,3	A				
	9.01	165 ± 0,8	177 ± 0,8	174 ± 0,7	A				
	10.00	156 ± 0,7	175 ± 0,8	175 ± 0,8	A				
	11.00	163 ± 0,7	181 ± 1,2	182 ± 0,6	A	163 ± 3,1	181 ± 2,7	182 ± 2,8	B
	12.00	166 ± 1,5	182 ± 1,5	180 ± 2,0	A	164 ± 3,2	181 ± 2,4	193 ± 1,9	B
	13.00	168 ± 1,5	179 ± 1,7	182 ± 1,5	A	183 ± 2,6	197 ± 1,6	207 ± 3,1	B
	14.00	165 ± 1,6	177 ± 1,9	177 ± 1,2	A	158 ± 6,9	182 ± 6,6	87 ± 7,2	B
	15.00	168 ± 2,4	180 ± 2,0	177 ± 1,9	A	172 ± 2,7	186 ± 2,3	93 ± 3,8	B
	15.06	165 ± 1,9	187 ± 1,7	185 ± 1,8	B				
	16.00	170 ± 2,0	185 ± 3,4	183 ± 2,3	A	172 ± 2,0	184 ± 1,3	95 ± 3,2	B
	16.08	169 ± 3,2	183 ± 1,5	186 ± 1,5	B				
	17.00	167 ± 1,4	182 ± 1,1	180 ± 2,0	A	172 ± 4,9	197 ± 2,2	205 ± 2,7	B
	17.10	165 ± 1,6	182 ± 2,3	186 ± 1,4	B				
	18.02	173 ± 3,2	185 ± 1,9	182 ± 2,1	A	174 ± 2,0	194 ± 2,0	202 ± 2,2	B
	18.10	167 ± 1,7	186 ± 2,2	185 ± 1,7	B				
	19.02	171 ± 2,7	186 ± 2,0	186 ± 1,9	A	166 ± 3,7	195 ± 4,5	202 ± 4,6	B
26.8.	5.12	187 ± 2,7	197 ± 1,7	202 ± 1,2	A	178 ± 2,8	205 ± 4,9	204 ± 3,1	B

Die erneuten Feinbewegungsmessungen am Gurgler-Ferner sollten zur Klärung folgender Fragen beitragen :

a) Es wurde vermutet, daß die auffallend geringe Fließgeschwindigkeit von ca. 2 cm/Tag, wie sie im Vorjahr ermittelt worden war, vom zufälligen Aufstellungsort der Marken abhing.

b) Um die Meßgenauigkeit noch weiter zu steigern, sollte neben einem Wild T2 auch ein Wild T3 verwendet werden.

Die Messungen konnten in der Zeit vom 22.-26.8.56 dank der Unterstützung durch den Deutschen Alpenverein durchgeführt werden. Als Pfeiler für die Instrumente dienten die massiven Fensterbänke des Zöllhäuschens beim Hochwildehaus.

In der Entfernung $e = 530$ m ($k = 0{,}83^{mm}/cc$) waren wiederum die drei Eismarken : Kegel = I, Stab = II, und Brett = III in Dreiecksform von 56 bis 90 cm Seitenlänge aufgestellt worden. Die « künstliche Moräne » um den Stab herum wurde noch stärker als im Vorjahr ausgelegt. Die Mire stand genau an derselben Stelle wie 1955.

Durch zeitweilige heftige Schneestürme, Regenschauer und Nebel sowie andere Umstände, waren die Messungen sehr stark behindert worden, so daß sie nicht lückenlos vonstatten gehen konnten. Es gelang jedoch, mit dem T3 und zwei Marken 1,5 Tage lang im stündlichen Rhythmus durch *einen Beobachter* zu messen. An der dritten Marke fiel für 6 Stunden die Beleuchtung aus. Darüberhinaus gelangen vor- und nahher noch einige Ablesungen in unregelmäßigen Abständen, so daß insgesamt eine Zeitspanne von 3,5 Tagen überbrückt werden konnte.

Als Meßfolge war folgendes Verfahren gewählt worden :

Mire — Stab — Brett — Kegel — Mire,

wobei jedes Ziel für sich dreimal nacheinander eingestellt und abgelesen wurde. Die Mirre wurde deshalb am Anfang und Ende jeder Reihe eingemessen, weil ein Fehler in der Visur dorthin in die Ergebnisse aller drei Marken eingeht, und fälschlich eine gemeinsame Bewegung vortäuscht. Es blieb problematisch, wie man bei größeren Differenzen zwischen Anfangs- und Endablesung zur Mire die Winkel m_γ bilden sollte. Hier wurde auch in diesen Fällen das Gesamtmittel aller sechs Ablesungen verwendet.

Die Fehlerrechnung lieferte folgende Ergebnisse :

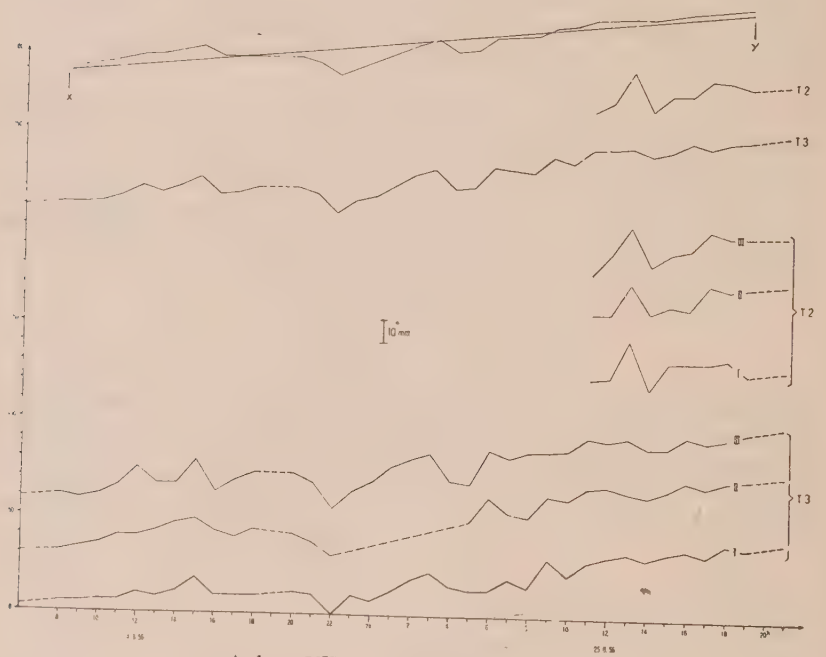
TABELLE 10

Ergebnisse der Fehlerrechnung

Stand	Mire			Kegel		Stab		Brett	
	m_1	m_2	m_r	m_r	m_γ	m_r	m_γ	m_r	m_γ
	cc	cc	cc	cc	cc	cc	cc	cc	cc
T3	± 1,3	± 1,1	± 1,1	± 1,2	± 1,7	± 1,2	± 1,6	± 0,9	± 1,5
T2			± 2,6	± 2,3	± 3,4	± 1,7	± 3,0	± 2,2	± 3,5

Die Fehler m_1 beziehen sich auf das erste und m_2 auf das zweite Beobachtungs-Tripel zur Mire. Wenn diese 6 Beobachtungen einer homogenen Meßreihe zugehören würden, müßte m_r kleiner sein als m_1 und m_2 und theoretisch etwa $\pm$ 0,8cc betragen. Der tatsächliche Wert $m_r = \pm$ 1,1cc enthält also auch den Einfluß der « Pfeilerdrehung ». Von der absoluten Sicherheit der Visuren zur Mire hängt das Ergebnis einer ganzen Untersuchung ab. Es hat sich gezeigt, daß bei ungünstigen Sicht- und Beleuchtungsverhältnissen die Mire nicht immer gut einstellbar war. Es ist daher vorgesehen den Durchmesser der Zielscheibe für die Mire von 10 cm auf 20 cm zu vergrößern. Ihre unveränderliche Aufstellung muß durch genaue Beobachtung (in mehreren Sätzen) des Winkels zwischen der Mire und einem guten Hochziel vor Beginn und nach Beendigung eines Unternehmens kontrolliert werden.

In der Anlage VII ist der Teil, welcher zusammenhängend in stündlichen Messungen ermittelt wurde, graphisch dargestellt. Im Vergleich mit den Ergebnissen



Anlage VII — Gurgler-Ferner 1956.

der Jahre vorher läßt sich eine Steigerung der Parallelität der drei Marken-Bewegungen feststellen. Für die restlichen Unregelmäßigkeiten bleibt die Frage offen: Sind es

a) Eigenbewegungen der Marken
b) Detailbewegungen der Eispartien oder
c) kurzfristige Ruckbewegungen?

Bei zukünftigen Messungen gilt es zunächst einmal den Fall a) auszuschließen, indem Marken verwendet werden, deren Stab nicht aus Holz besteht, sondern aus einem weißen Kunststoffrohr, in dessen Inneren eine Kältemischung (Eis-Salz o.ä.) das Festfrieren der Marke bewirken soll. Ein noch zu erprobender Strahlungsschutz muß dann das erneute Lockerschmelzen verhindern.

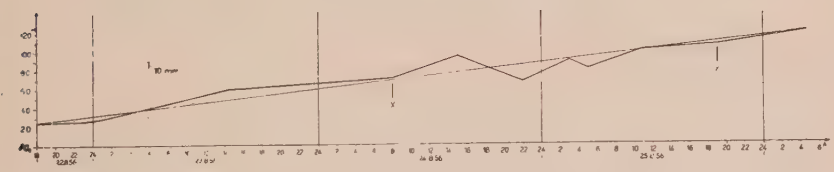
Erst wenn a) unschädlich gemacht ist, kann mit Hilfe flächenhaft verteilter Marken und zweier Theodolite, an welchen streng gleichzeitig aber in gegenläufiger Reihenfolge beobachtet wird, Fall b) von c) getrennt werden.

So lange noch so große Differenzen zwischen den Messungen mit T3 und denjenigen mit T2 zu verzeichnen sind, wie sie Anlage VII für die Zeit 25.8., 11-19h demonstriert, kann noch nicht von eindeutigen Ergebnissen gesprochen werden. Als Hauptursache dürfte wohl die bereits erwähnte unsichere Mireneinstellung angenommen werden.

Zuerst müssen diese Unterschiede aus einer Meßreihe verschwunden sein, ehe es Zweck hat, die zweidimensionale Bestimmung des Bewegungsablaufes in Angriff zu nehmen, die zeigen kann, ob die Pulsationen in Wirklichkeit einer flächenhaften Turbulenz entsprechen. Beim Vorwärtseinschneiden der Marken von einer Standlinie aus müssen die bisher noch nicht getilgten Fehler zu einem starken seitlichen Hin- und Herspringen der gefundenen Markenorte führen, denn der meist unvermeidliche spitze Schnittwinkel der Visuren bewirkt, daß sich die Fehler mit dem Cosecans dieses Winkels vervielfachen. Einige Versuchsmessungen 1955 und 1956 haben diese Tatsache bestätigt.

Der obere Teil der Anlage VII gibt noch die Mittelkurven T3 und T2 sowie die Glättungskurve X-Y für die T3-Beobachtungen. Ein Vergleich (zum Beispiel) zwischen dieser und T3 II zeigt, daß sich die zufälligen Meßfehler nur noch gering auswirken.

Um nun noch den Bewegungsablauf über den ganzen Beobachtungszeitraum zu zeigen, sind in Anlage VIII auch die Steumessungen aufgetragen. Der Abschnit



Anlage VIII — Gurgler-Ferner 1956.

X bis Y ist dabei (etwas generalisiert) der Anlage VII entnommen. Wiederum ergibt sich das Bild von « langatmigen » unregelmäßigen Pulsationen mit Amplituden von etwa 10 mm.

Für die Station 1956 hat sich nur eine geringfügig größere Durchschnittsgeschwindigkeit

von $V\ 1956 = 24^{mm}/Tag$

gegenüber $V\ 1955 = 18{,}7^{mm}/Tag$

ergeben.

Wenn trotz erwähnter Schwierigkeiten viele Probleme gelöst werden konnten, so war dies vor allem den Mitarbeitern zu danken, die an den oft strapaziösen Tag- und Nachtmessungen mitgewirkt haben. Weitere Untersuchungen werden folgen, um die angedeuteten neuen Lösungen zu erproben.

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ON THE DEFORMATION OF EXCAVATIONS IN THE GREENLAND NÉVÉ

JOSEPH K. LANDAUER
U. S. Army Snow Ice and Permafrost Research Establishment Corps of Engineers, U. S. Army Wilmette, Illinois

Abstract

Deformations occurring in a tunnel, two trenches, and a 30 m deep pit excavated in the Greenland névé in the summer of 1954 have been measured over a period of 2 years.

The experimental results indicate that closure rates increase with lateral distance from a restraining boundary up to distances of about 1 m. At larger distances, the effect of the boundary is not appreciable. Closure rates for deep excavations are not found to be strongly depth-dependent. This is due to a roughly parallel increase of viscosity and pressure with depth. Vertical compaction results from the pit agree well with those calculated from the depth-density relation using Sorge's Law.

A theoretical calculation for the tunnel and pit closure, which is a modification of existing theories for the deformation of an elastic, compressible, thick-walled cylinder, is in fair agreement with the observed deformations. From this agreement it appears that the flow behavior of the névé can be described on the basis of a Newtonian viscosity.

The trench closure is described as the squeezing out of horizontal layers, fixed at some distance from the trench, and modified only slightly by the drag of neighboring layers.

Measurements will be continued during future summers.

Introduction

The deformations of excavations in the Greenland Ice Cap are of interest in that they provide information on the lifetime of installations and, of more fundamental importance, they lead to an understanding of the plastic nature of the ice cap.

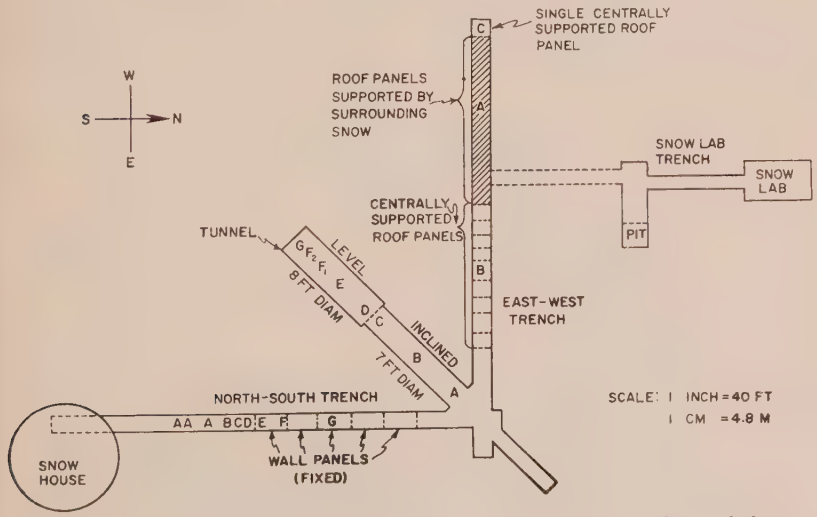


Fig. 1 — Location of deformation measurements at experimental site.

TABLE I

Tunnel closure

Location			Closure rate (cm/yr), strain rate (yr^{-1}) for 2nd year							
Distance from end wall (m)		Depth of Center below Aug. 54 surface (m)	Position 1		2		4		3	
			(cm/yr)	yr^{-1}	(cm/yr)	yr^{-1}	(cm/yr)	yr^{-1}	(cm/yr)	yr^{-1}
Inclined										
WA	18.3	3.7	—	—	—	—	5.2	0.024	35.7	0.23
WB	14.2	4.9	<0	<0	7.7	0.040	6.4	0.033	25.0	0.17
WC	10.5	6.0	0	0	6.6	0.034	5.4	0.027	22.5	0.14
Horizontal										
WD	8.8	6.1	—1.8	—0.0075	6.1	0.026	7.0	0.031	21.6	0.10
WE	3.9	6.1	—0.7	—0.0027	7.8	0.033	7.8	0.033	22.2	0.11
WF_1	2.1	6.1	—0.1	—0.0004	6.6	0.026	7.5	0.030	21.0	0.11
WF_2	1.2	6.1	—0.2	—0.0008	7.1	0.028	6.4	0.025	18.6	0.096
WG	0.3	6.1	+0.2	+0.0009	5.4	0.022	4.4	0.018	14.6	0.070
End wall	0	6.1		+~0.003						0.051

The tunnels, trenches, and pit under study were excavated in the summer of 1954 at a site about 200 miles east of Thule, Greenland, at an elevation of about 6800 ft above sea level. Details of the excavations can be found in SIPRE Report 20 (Snow, Ice and Permafrost Research Establishment, 1955). A diagram of the experimental site is shown in Figure 1. The instrumentation was installed during the summer of 1954 and readings were made biweekly for one year. After the summer of 1955 no measurements were made until the summer of 1956. More readings will be taken during future summers. Two years of data are reported on here.

Deformation measurements were made at various locations in order to find the effect of differing boundary conditions. Eight locations were chosen in the tunnel to obtain the closure rate as a function of distance from the end. In the north-south trench, closure rate was determined at five distances from lateral restraining panels, and in the east-west trench values were found under three types of roof-loading conditions. Location of the measurements in the tunnel and trenches is shown in Figure 1. The deep pit was instrumented for measurement of lateral closure and vertical compaction.

Measurements were of two types. The closure rates were determined by measuring the separation between 1/4-in. wood dowels driven about 20 cm into the snow. A sliding rule gave values of the separation reproducible to $\pm$ 1 mm. Compaction was determined by a wire and dial arrangement which indicated the separation of two pegs. Recorder wire was run from one peg around a bearing on the second peg. Changing the distance between pegs caused a rotation of the bearing which could be read to $\pm$ 1 deg. This gives an accuracy of $\pm$ 0.1 mm in the compaction values. A wire from the bottom to the top of the deep pit was used to measure the total compaction of the pit with an accuracy of several millimeters.

Measurements and Results

Tunnel

The tunnel was approximately cylindrical and consisted of two sections. An inclined section, 7 ft in diameter and 30 ft long, led to an 8 ft diam, 30-ft long, horizontal section, the center of which was 20 ft below the August 1954 surface (see Fig. 1). Four peg pairs were placed on the horizontal, vertical and 45 deg diameters at eight locations. The coding scheme for these measurements is shown in Figure 2*a*.

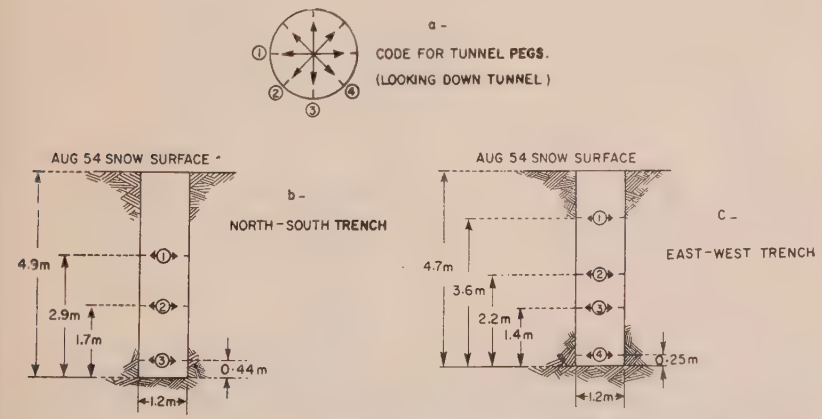


Fig. 2 — Trench and tunnel measurements. August 1954 dimensions are given.

The values of most fundamental interest are in the horizontal sections, as the inclined section was influenced by the adjacent trenches. Closure was determined at several distances from the end wall (see Table 1). Deformation curves as a function

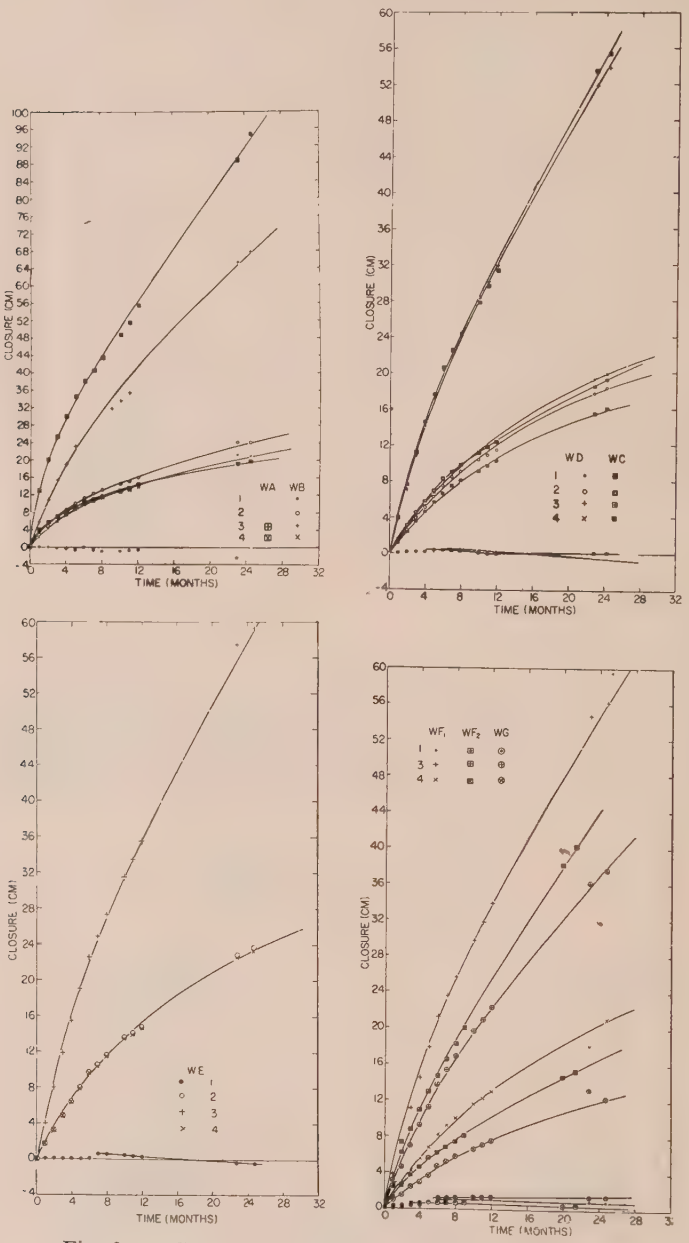


Fig. 3 — Tunnel deformation (3*a*, 3*b*, 3*c*, and 3*d*).

of time are given in Figure 3. During the first year, measurements were made about once a week. In order to facilitate plotting, these are shown by the month. Only slightly greater scatter is observed on the complete plots. In Table I the closure rates for the second year are given. The strain rates based on these values have been obtained by dividing the second year closure rates by the average diameter of the tunnel at the corresponding points.

The end wall of the tunnel was instrumented with vertical and horizontal dial sets. The horizontal dial stuck occasionally and gave reliable data only for short time intervals. Therefore, the value indicated may be inaccurate. The vertical diameter gave consistent values for the entire 2-yr period.

In August 1956 density profiles were taken near section E in the tunnel. Vertical cores were taken in the floor and roof and a horizontal core at the level of peg pair E1. The density values are shown in Figure 4, where they are compared with the density expected if no tunnel were present.

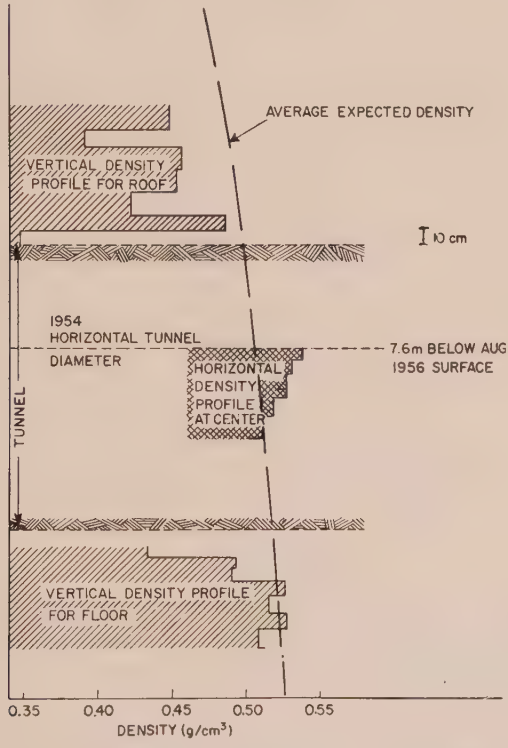
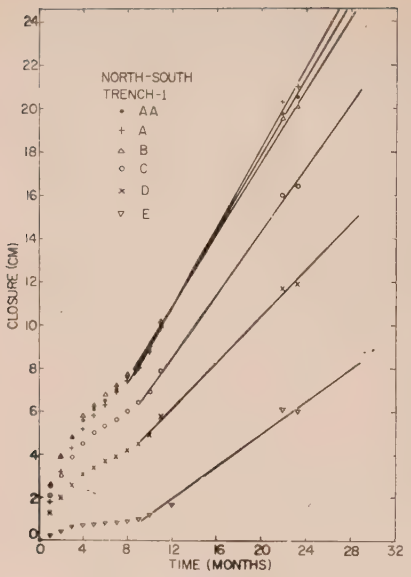
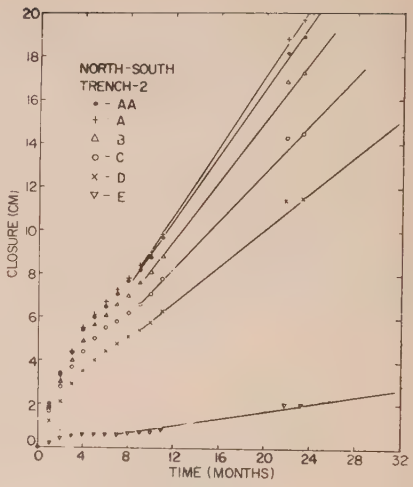


Fig. 4 — Tunnel density profile near section E, for August 1956. Bold dashed line shows average density to be expected if no tunnel were present.

North-south trench

Three tiers of restraining panels were installed along part of the north-south trench wall. These were arranged so that horizontal closure was restricted but vertical motion was nearly unrestrained. In this way the panels were more or less free to settle vertically as the snow compacted. Three horizontal rows of pegs were used to measure





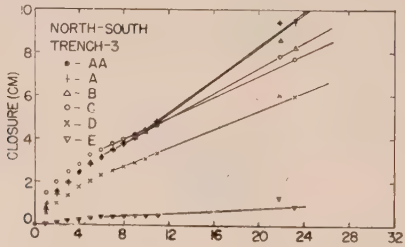


Fig. 5 — North-south trench deformation. (*5a*, *5b*, and *5c*).

TABLE II
North-south trench closure

Location	Horizontal distance from fixed panels (m)	Closure rate for second year (cm/yr) Position 1	2	3
AA	3.66	10.6	9.1	4.6
A	2.44	11.2	9.6	4.6
B	1.22	10.1	8.3	3.5
C	0.61	8.8	6.6	3.0
D	0.30	6.5	5.1	2.6
E	South end of end panel	4.5	1.1	0.4
F	North end of end panel	3.3	3.0	2.8
G	Center of central panel	3.3	2.8	2.7

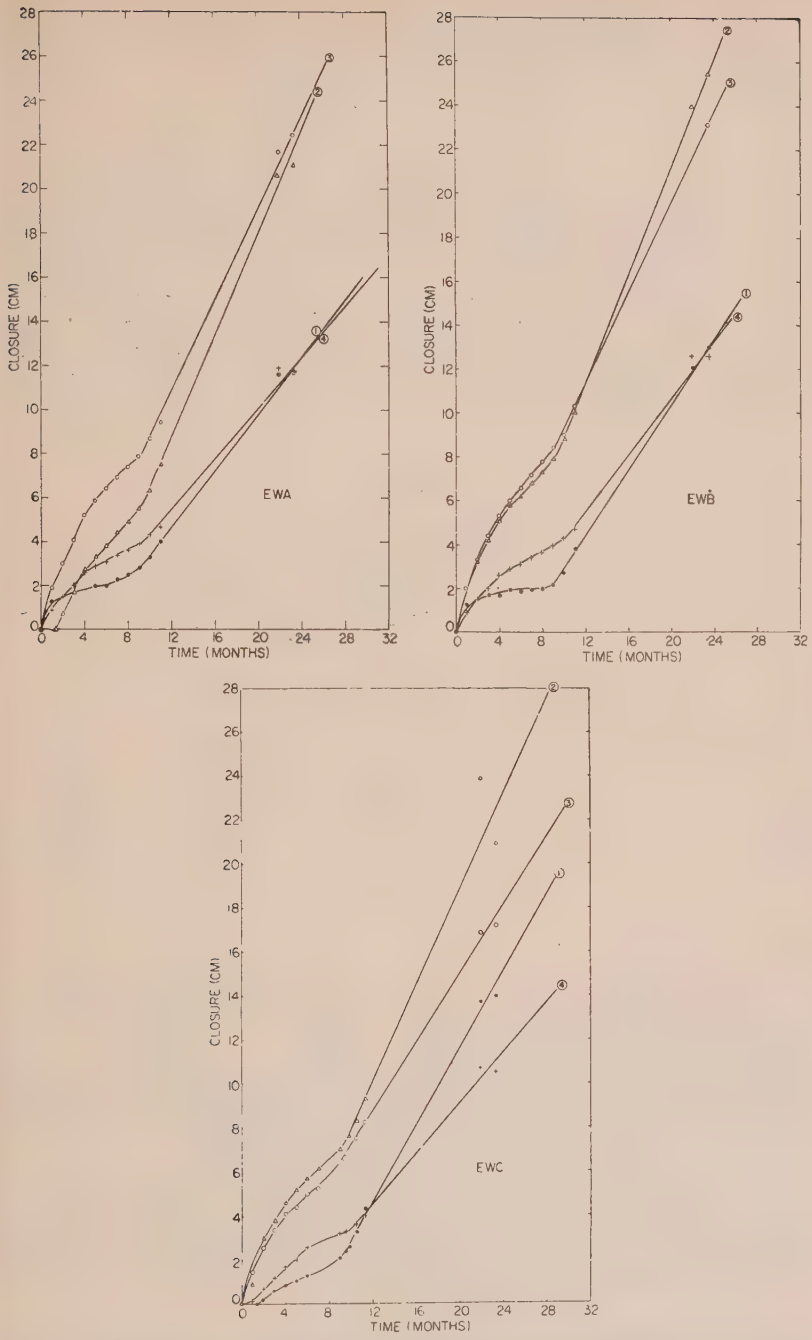


Fig. 6 — East-west trench deformation. Numbers on curves indicate position as in Figure 2*c*. (6*a*, 6*b*, 6*c*).

the closure as a function of distance from the panels. The code for these pegs is shown in Figure 2*b*, and the data is plotted in Figure 5. Closure rates for the second years are shown in Table II. It can be seen that the restraining panels were not absolutely rigid but allowed some small closure.

East-west trench

The effect of three different roof-supporting conditions on the closure of the east-west trench was determined. In section A, the roof panels were laid directly on the snow. Closure rates were measured below the center of this section. In section B, roof panels were supported by rigid columns resting on the trench floor. Measurements were obtained at the center of this 14-m section of supported roofing. At the west end of the trench, a single column supported a 4-ft diam. circular plate, forming section C. Four pairs of pegs were placed in hard snow layers at depths as shown in Figure 2C. The closure data is shown in Figure 6 and closure rates for the second year given in Table III.

TABLE III

East-west trench closure

Location	Second year closure rates (cm/yr) Position			
	1	2	3	4
Initial depth (m)	1.3	2.6	3.4	4.6
1956 depth (m)	3.3	4.5	5.3	6.3
A	7.7	13.8	12.3	6.9
B	8.9	14.8	12.4	7.8
C	10.5	13.0	9.3	6.8

Deep pit

A 30.5-m deep pit, about 2 m square in cross-section, was instrumented for both horizontal closure and vertical compaction. The closure was measured at seven levels (Table IV). Closure data is shown in Figure 7.

Vertical compaction between selected levels was measured with the dial sets. Because of sticking (and the subsequent need for resetting), these results were obtained only for the first year. The strain rates for December 1954 and June 1955 are given in Table V. The June 1955 strain rates are plotted in Figure 8.

The overall vertical compaction was measured by a long wire fixed at the pit bottom and running over a pulley at the top. The measured velocities were 64 cm/yr as of December 1954; 48 cm/yr as of June 1955; and 49 cm/yr as of May 1956. Because the pit is inclined 15 deg. from the vertical, it follows that the vertical compactive velocities are as shown in Table VI.

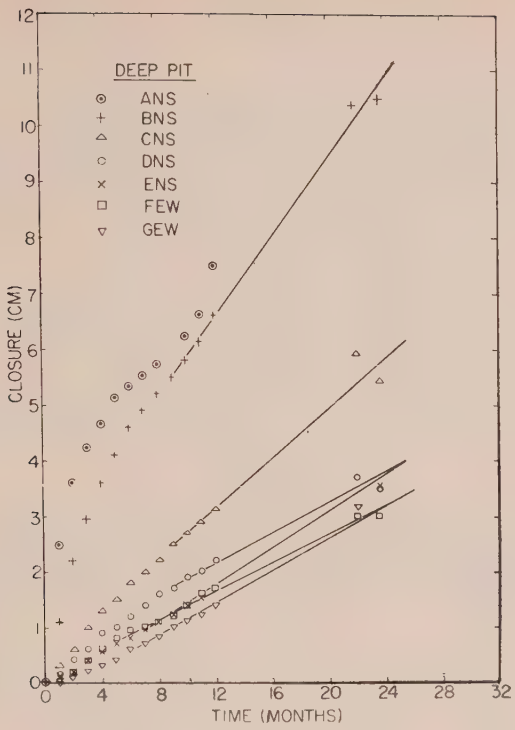


Fig. 7 — Deep pit closure.

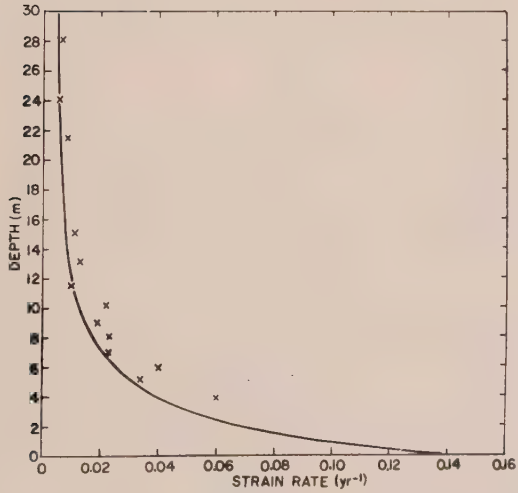


Fig. 8 — Vertical strain rates in the deep pit for June 1955. The solid curve is plotted from Eq. 3 and the observed depth-density relation.

TABLE IV

Deep pit closure rates.

Position	Depth below snow surface July 54, measured (m)	Depth below snow surface Aug. 56 calculated (m)	Second year closure rate (cm/yr)	Strain rate for second year (yr^{-1})
ANS	1.04	3.11	—	—
BNS	3.11	5.06	4.3	0.023
CNS	6.40	8.05	2.7	0.013
DNS	10.50	12.08	1.65	0.008
ENS	14.50	16.01	2.0	0.010
FEW	20.50	21.98	1.5	0.009
GEW	29.00	30.42	1.8	0.012

TABLE V

Vertical compaction in deep pit

Position	Initial depth of layer below July 54 surface (m)	Calculated depth of center of layer June 55 (m)	Strain Rate Dec 54, measured (yr^{-1})	Strain Rate June 55, measured (yr^{-1})	Strain Rate Calculated from Sorge's Law using Eq. 3 (yr^{-1})
P 1	2.1-3.8	3.9	0.085	0.060	0.052
P 2	3.8-4.7	5.2	0.053	0.034	0.039
P 3	4.7-5.5	6.0	0.059	0.040	0.032
P 4	5.5-6.7	7.0	0.033	0.023	0.026
P 5	6.7-7.8	8.1	0.032	0.023	0.020
P 6	7.8-8.5	9.0	0.022	0.019	0.018
P 7	8.5-10.3	10.2	—	0.022	0.015
P 8	10.3-11.3	11.6	—	0.010	0.012
P 9	11.3-13.6	13.2	0.013	0.013	0.011
P 10	13.6-15.2	15.1	—	0.011	0.009
P 11	19.7-21.9	21.5	0.009	0.0086	0.006
P 12	22.7-24.2	24.1	0.008	0.0054	0.006
P 13	26.6-28.2	28.1	0.007	0.0067	0.005

TABLE VI

Overall pit compaction

Date	Vertical compactive velocity (cm/yr)	Strain rate (yr^{-1})
Dec 54	66	0.022
June 55	49.5	0.017
May 56	50.5	0.017

Sorge's Law

At the location of the test site, melting is a very infrequent phenomenon and the snow can be considered dry. There is good reason to believe that no significant climatic changes have occurred over the period of its accumulation. As reported in SIPRE Report 20 (p. 14, 32), the average accumulation, A, is 41.5 g/cm²-yr and the temperature of the snow pack at depths below 10 m is $T = -24.4$ C. The depth-density relation for the upper 50 cm can be expressed approximately as:

$$\gamma = 0.492 + 5.26 \times 10^{-3}z - 0.152\, e^{-0.217z} \tag{1}$$

where γ is the density (g/cm³) and z the depth below the surface (m). This relation implies that the density at a particular depth is independent of time, which is a statement of Sorge's Law (Bader, 1954). The pressure for any depth to 50 m can be calculated fairly accurately by integrating Eq. 1. The depth-time of deposition curve can be obtained from the relation:

$$\sigma_z = At = \int_0^z \gamma dz \tag{2}$$

where σ_z is the vertical pressure (g/cm²) at a level deposited t yr ago.

Levels originally measured below the 1954 surface were deeper in 1955 and 1956. The velocity, dz/dt, of any point relative to the surface is given by:

$$dz/dt = A/\gamma$$

The vertical strain rate, $\dot{\varepsilon}_z$, derived from the above relation is:

$$\dot{\varepsilon}_z = \frac{d}{dz}\left(\frac{dz}{dt}\right) = -\frac{A}{\gamma^2}\left(\frac{d\gamma}{dz}\right), \tag{3}$$

$\dot{\varepsilon}_z$ is plotted as the solid curve in Figure 8.

Equations 1 through 3 are consequences of an equilibrium condition on the ice cap. Evidence supporting this will not be presented here; Sorge's Law will be taken as an assumption.

Deformation of a cylindrical tunnel in a viscous compressible medium

Nye (Nye, 1953) has treated the closure of tunnels in ice on the basis of flow of an incompressible material whose strain rate varies as a power of the stress. Snow, being compressible, must be treated differently. Further, for simplicity, a linear flow law is used. (It will be shown later that this assumption leads to adequate agreement between theory and experiment.) Assuming, then, that snow deforms by Newtonian flow, the deformation of a cylindrical tunnel can be calculated from the equations already developed for long, thick-walled, compressible elastic tubes by replacing the elastic parameters with viscous ones. An additional assumption is that the density and pressure of the snow are constant in the region where the stress distribution created by the tunnel is not negligible. As usual, it is also assumed that the snow is homogeneous and isotropic, and end effects are neglected.

The solution for the elastic case involves a radial stress, σ_r, which decreases from the value p at the inner boundary to zero at the outer boundary. For the tunnel, the radial stress must be zero at the inner wall and equal to a constant compressive value, p, at sufficiently large distances. To obtain this condition the deformation for an internal radial stress, $-p$, is calculated using the modified elastic equations, and

to this is added the deformation resulting from a constant stress, p, not varying with radial distance. This additional stress contributes a component which is the same as if no tunnel were present.

According to Hill (1950, p. 108) the radial displacement, u, in a long thick-walled cylindrical tube of inner radius a and outer radius b, for a compressible elastic medium is:

$$u = -\nu\varepsilon_l r + \frac{(1+\nu)\,p}{E\,(b^2/a^2-1)}\,[(1-2\nu)\,r + b^2/r] \qquad (4)$$

where r is the radial distance from the tube axis, ν is Poisson's ratio, E is Young's modulus, and p is the internal pressure. The tube axis is in the l direction and ε_l is the strain in that direction. If b is taken very much larger than a, and u is calculated at distances small compared to b, Eq. 4 reduces to:

$$u/r = -\nu\varepsilon_l + (1+\nu)\,pa^2/Er^2 \qquad (5)$$

The change to viscous parameters is effected by replacing u/r by the strain rate $\dot{\varepsilon}_r$; ε_l by the longitudinal strain rate, $\dot{\varepsilon}_l$; and E by the viscosity, η, multiplied by $2\,(1+\nu)$ (Love, 1926, p. 104). The plastic analogue of Poissons's ratio, m, defined as the negative of the ratio of the uniaxial strain rate in the direction of the stress to that normal to it (Sipre Report 7) replaces $1/\nu$ Therefore, for a viscous, compressible tube with infinite outer radius, the radial strain rate under internal pressu.e will be:

$$\dot{\varepsilon}_r = -\dot{\varepsilon}_l/m + a^2p/2\eta r^2 \qquad (6)$$

At the tube wall, $r = a$, and Eq. 6 becomes:

$$-\dot{\varepsilon}_a = \dot{\varepsilon}_l/m - p/2\eta \qquad (7)$$

Due to its essentially horizontal surface, the ice cap deforms only by settling and in the absence of a tunnel the only strain is $\dot{\varepsilon}_z$, the z direction being vertical. The vertical stress, σ_z, is given by Eq. 2. The horizontal stress, σ_y is such that the horizontal strain is zero. By analogy to elastic theory, the ratio of the two stresses is (Appendix Case II):

$$\sigma_z/\sigma_y = m - 1 \qquad (8)$$

A hydrostatic condition holds only for $m = 2$ (the value for ice); the deviation from hydrostatic is great for light snows.

At an angle, θ, to the vertical, the radial stress component is:

$$\sigma_r(\theta) = \sigma_z \cos^2\theta + \sigma_y \sin^2\theta$$

Using Eq. 8 this becomes:

$$\sigma_r(\theta) = [(m-2)\cos^2\theta + 1]\,\sigma_z/(m-1) \qquad (9)$$

$\sigma_r(\theta)$ is the stress which occurs when no tunnel is present or at a large distance from the tunnel. Therefore, the boundary conditions will be satisfied if the internal pressure is taken as:

$$p = -\sigma_r(\theta) \qquad (10)$$

The calculation of ε_l is based on the result obtained for the elastic thick-walled tube that the longitudinal stress, σ_l, is constant over the section (Hill, 1950, p. 107). Since the longitudinal stress is σ_y at a large distance from the tunnel, it follows that $\sigma_l = \sigma_y$. The longitudinal strain rate is, therefore, equal to zero.

The viscosity is determined from the rate of compaction of the snow in the absence of a tunnel. A correction to the observed ratio of vertical stress to strain must be made to compensate for the confined condition (Appendix, Case II). Hence the viscosity is:

$$\eta = (m - 2)\,\sigma_z/2\,(m - 1)\,\dot{\varepsilon}_z \tag{11}$$

In Table VII values of depth, density, σ_z, and η are given. The viscosity as a function of depth is given in Figure 9.

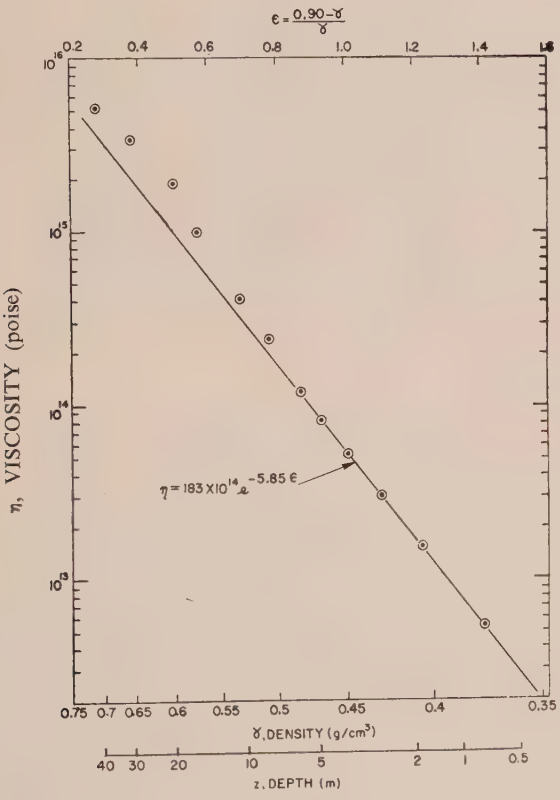


Fig. 9 — Viscosity calculated from Sorge's Law using observed depth-density values and corrected for lateral constraint.

The radial deformation component due to the constant stress term, $\dot{\varepsilon}_z \cos\theta$ must be added to Eq. 7. Using Eqs. 7, 9, 10, and 11 (for $\dot{\varepsilon}_z$), the final expression for the tunnel deformation becomes:

$$-\dot{\varepsilon}_a = \frac{(m - 2)\,\sigma_z}{2\,(m - 1)\,\eta}\left(\cos^2\theta + \cos\theta + \frac{1}{m - 2}\right) \tag{12}$$

TABLE VII

Average parameters for Site 2.

z, depth below snow surface (m)	γ (g/cm^3)	σ_z, calc. * (kg/cm^2)	m, exp.†	η, calc. (poise)
0	0.340	0	7.4	—
1	0.375	0.037	6.8	0.52×10^{14}
2	0.405	0.078	6.4	0.15
3	0.429	0.118	6.1	0.30
4	0.449	0.160	5.8	0.52
5	0.468	0.203	5.6	0.82
6	0.482	0.250	5.4	1.2
8	0.507	0.350	5.2	2.4
10	0.533	0.452	5.0	4.1
15	0.574	0.730	4.6	10
20	0.600	1.016	4.3	19
30	0.653	1.641	4.0	34
40	0.706	2.318	3.6	52

*From Eq. 2.
†From SIPRE Report 7.

DISCUSSION

Tunnel

Referring to Figure 3 and Table I, the general tendencies evident in the tunnel deformation are:

1. Closure rates for the two oblique diameters do not show any trend in their differences except in the regions influenced by near by trenches. In the uninfluenced region, the two can be considered the same. The assumption of horizontal isotropy is strengthened by this observation.

2. The end wall restrains the closure for only a short distance. At a distance of 1 to 2 m the closure rates reach values that do not increase for greater distances.

3. The compactive velocity on the end wall is somewhat greater than that determined at the same level in the deep pit. This can be explained by the stress concentration on the end wall of the tunnel.

4. Closure along the vertical diameter is about three times faster than along the oblique diameters.

5. Closure rates for the horizontal diameters are close to zero with some values being negative.

The closure theory presented in the preceding section can be applied to the tunnel As an example, for position E of the tunnel, as of the summer of 1956 the center was 7.6 m below the surface. Reference to Table VII provides the parameters at 7.6 m for the solution of Eq. 12. The result is:

$$-\dot{\varepsilon}_a = 6 \times 10^{-10} (\cos^2\theta + \cos\theta + 0.3) \text{ sec}^{-1}.$$

The values for $\theta = 0°$, 45°, and 90° are:

Position	θ	$-\dot{\varepsilon}_a$ theoretical (sec^{-1})	$-\dot{\varepsilon}_a$ observed at E (sec^{-1})
3	0	14×10^{-10}	35×10^{-10}
2,4	45	9	10
1	90	2	—0.9

Certain of the assumptions made are probably responsible for the discrepancy between observed and calculated results. The use of the thick-walled cylinder theory for tunnel deformation implies that the tunnel is away from the influences of the surface and that the medium is isotropic and homogeneous. The ratio of depth below the surface to tunnel radius is actually about 8 to 1. Since the stress drops off as a^2/r^2, the first assumption should be adequate. Measurements on the physical properties of the snow indicate that anisotropy exists for small samples but is not too great. The overall condition is entirely different in that the density varies in the vertical but not in the horizontal direction. The viscosity over the tunnel varies from 2.2×10^{14} to less than 0.05×10^{14} poise at the surface. Because of the lower viscosities above, actual vertical closure will be greater than the value calculated using the viscosity value at the center of the tunnel.

Trenches

Figures 5, 6 and Tables II, III show that the trenches close at a greater rate near the center than at the roof or floor. Restraint by these more or less rigid boundaries reduces the motion from the unrestrained value.

Over the two years of study no significant differences appear between the three boundary conditions on the roof sections of the east-west trench. This independence may not hold in the future when the overburden of snow has increased.

In the north-south trench, lateral constraint reduces the closure at distances up to about 1 m. This is consistent with the observations on the constraint produced by the end wall of the tunnel.

Closure of the trenches apparently takes place by lateral expansion of the horizontal layers, modified to some extent by the drag exerted by neighboring layers. Each layer remains fixed at some distance from the trench but is free to move at the trench wall. The relative amount of movement, $\dot{\varepsilon}_y$, is found for the semi-compactive case (Appendix, Case III) to be $-\dot{\varepsilon}_y = \sigma_z/2m\eta$. The observed trench closure rate, V, must be normalized by some distance, d, to give a strain rate that can be compared with $\dot{\varepsilon}_y$. The lateral strain rate is then :

$$-\dot{\varepsilon}_y = V/2d. \tag{13}$$

d is of the order of the effective distance of disturbance due to the trench. Note that the trench closure is independent of its width. From Table III, the closure rate in the east-west trench is about 14 cm/yr at 4.5 m depth and about 12 cm/yr at 5.3 m. The calculated values of $\dot{\varepsilon}_y$, interpolated from values in Table VII, are 2.5×10^{-10} sec^{-1} and 2.2×10^{-10} sec^{-1} at these depths. From Eq. 13, the two values of d are 9.0 m and 8.5 m, respectively.

The trench deformation concept presented above, in which the deformation rate of a horizontal layer is roughly independent of the motion of its neighbors, leads to a somewhat large value for d. Closure measurements on deeper trenches should be made in order to verify Eq. 13.

Deep pit

Lateral closure of the pit is roughly constant with depth (see Table IV). These results are understandable; although the load increases with depth, the viscosity also increases. Calculation of the closure rate can be made using the tunnel theory. For this application the square cross section of the pit is assumed circular with its diameter equal to the side of the square. In Eq. 7 the longitudinal strain rate for the tunnel, $\dot{\varepsilon}_l$, is replaced by:

$$\dot{\varepsilon}_z = (m - 2)\,\sigma_z/2\eta\ (m - 1)$$

(Appendix, Case II), and the normal stress at the tunnel wall, p, is replaced by $-\sigma_y$. The constant radial stress, p, is also replaced by σ_y and, therefore, contributes no deformation. Equation 7 becomes, using Eq. 8,

$$-\dot{\varepsilon}_a = \frac{\sigma_z}{m\eta}. \tag{14}$$

Comparison of $-\dot{\varepsilon}_a$ calculated from the parameters in Table VII, with the observed values in Table IV shows:

Position	Observed closure rate (sec^{-1})	$-\dot{\varepsilon}_a$, Calculated (sec^{-1})
BNS	7.3×10^{-10}	4.3×10^{-10}
CNS	4.1	2.8
DNS	2.5	1.7
ENS	3.2	1.4
FEW	2.8	1.2
GEW	3.8	1.2

In converting the observed closure rates to strain rates the width of the pit was used as the equivalent of the diameter in the theoretical calculation. Therefore, the observed strain rates will be somewhat greater than the calculated ones, although this effect is not sufficient to account for the total difference between these values.

Observed and calculated vertical compactive velocities are compared in Table V. The 1955 overall strain rate for the pit of 0.017 yr^{-1} agrees well with the value 0.015 yr^{-1} calculated from Sorge's Law (Eqs. 1 and 3). It is understandable that the observed compactive strain rates are somewhat larger than the calculated values, for this calculation is based on parameters holding where no pit is present. Increased compaction should be expected due to the lessening of constraint at the pit walls. Local inhomogeneities and imperfect boundary conditions would be expected to limit, to some extent, the correspondence between observed and calculated results.

Mr Gunther Frankenstein of SIPRE deserves the credit for performing the bulk of the deformation measurements under the severe conditions during the 1954-55 winter.

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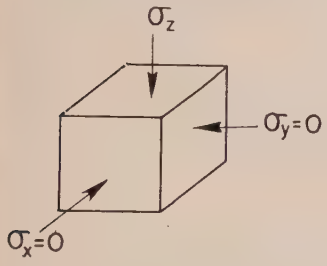
APPENDIX

By use of elastic theory, the principal stresses and strains for three different cases are calculated. This analysis is similar to that of Bucher (1956). Compressive stresses and strains are taken positive. The elastic results are modified for a compressible viscous material by substituting the strain rates for strains; $1/m$ for Poisson's ratio ν; and $2\eta\ (1 + \nu) = 2\eta\ (m + 1)/m$ for Young's modulus, E (Love, 1926, p. 104)

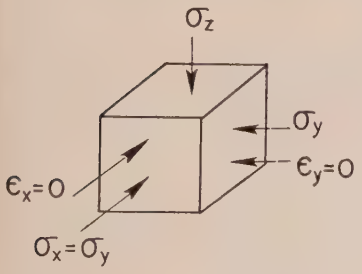
Case I. Uniaxial:



Elastic results:
$E\,\varepsilon_z = \sigma_z$
$E\,\varepsilon_x = E\,\varepsilon_y = -\nu\sigma_z.$

Viscous analogue:
$\dot{\varepsilon}_z = m\sigma_z/2(m + 1)\eta$
$-\dot{\varepsilon}_x = -\dot{\varepsilon}_y = \sigma_z/2(m + 1)\eta$

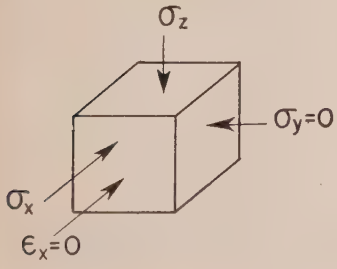
Case II. Compactive:



Elastic results:
$E\,\varepsilon_x = E\varepsilon_y = 0 = \sigma_x - \nu(\sigma_y + \sigma_z)$
$\therefore \sigma_x = \sigma_y = \nu\sigma_z/(1 + \nu)$
$E\varepsilon_z = \sigma_z - 2\nu\sigma_y = (1 - 2\nu)(1+\nu)\,\sigma_z/(1 - \nu).$

Viscous analogue:
$\sigma_z/\sigma_y = m - 1$
$\dot{\varepsilon}_z = (m - 2)\sigma_z/2(m - 1)\eta.$

Case III. Semi-Compactive:



Elastic results:
$E\varepsilon_x = 0 = \sigma_x - \nu\sigma_z \quad \therefore \sigma_x = \nu\sigma_z$
$E\varepsilon_y = -\nu(\sigma_x + \sigma_z) = -\nu(1 + \nu)\,\sigma_z.$

Viscous analogue:
$\sigma_z/\sigma_x = m$
$-\dot{\varepsilon}_y = \sigma_z/2m\eta.$

DRUCK- UND VERFORMUNGSMESSUNGEN IN EISSTOLLEN

R. HAEFELI
Gletscherkommission der Schweiz Naturforschenden Gesellschaft

RÉSUMÉ

La première partie de la présente étude donne et discute les résultats des mesures spéciales de la pression de la glace dans une galerie tempérée du glacier de Zmutt. La mesure des pressions exercées sur une colonne verticale, a été effectuée à l'aide d'un ressort. Le résultat le plus important, c'est que la pression maximale mesurée au bout d'un certain temps correspond à la charge de la glace sus-jacente. On peut déduire de ce fait certaines conclusions relatives aux forces en jeu lors de l'érosion glaciaire.

La seconde partie traite des nouvelles mesures de déformation dans la calotte glaciaire froide du Jungfraujoch. Ces mesures se rapportent d'une part à la déformation d'une galerie circulaire, d'autre part à la répartition des vitesses horizontales de rampement dans une section de la calotte.

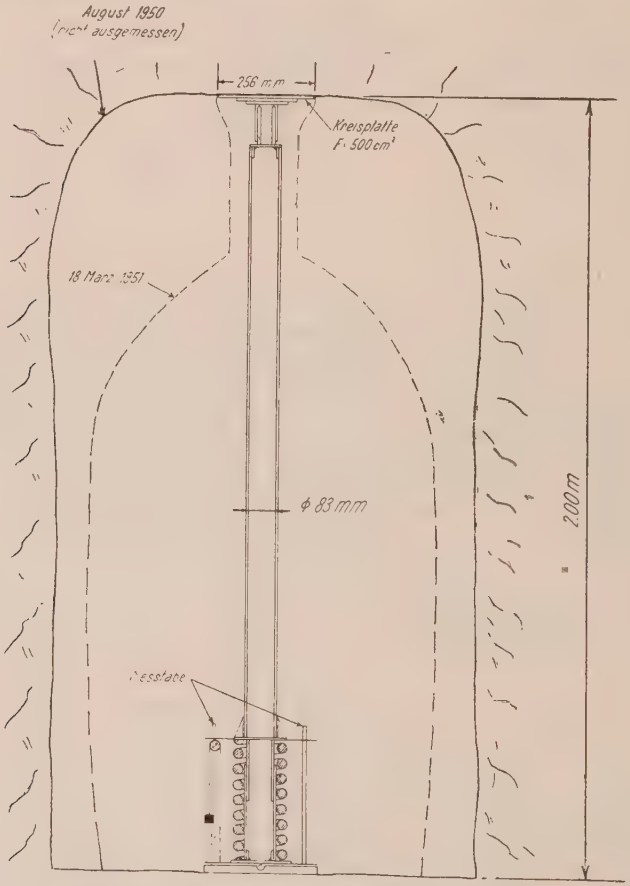


Fig. 1 — Querschnitt durch die Drucksäule des Z'Muttstollens (schematisch).

I. Eisdruckmessung in einem temperierten Gletscher.

In den Jahren 1948-1949 wurde in einem ca. 1000 m langen Eisstollen des Z'Muttgletschers der Versuch unternommen, den Eisdruck im temperierten Eis mit Hilfe einer einfachen Vorrichtung direkt zu messen. Zu diesem Zwecke wurde in einer Entfernung von 547 m vom Stollenausgang eine vertikale, elastische Drucksäule in eine Nische eingebaut, um den auf das obere Ende dieser Säule infolge der langsamen, plastischen Kontraktion des Stollens ausgeübten Druckes zu verfolgen. Die Eisüberlagerung betrug an der betreffenden Stelle, die sich in einer Druckzone des Gletschers befindet, 43 m. Ueber die in diesem Stollen ausgeführten Verformungs- und Geschwindigkeitsmessungen wurde an anderer Stelle berichtet [1].

1. *Versuchsanordnung*

Die Drucksäule bestand aus zwei teleskopartig ineinandergreifenden Metallrohren, zwischen denen eine Messfeder eingesetzt war (Fig. 1). Die kreisförmige Druckplatte am oberen Ende der Spannsäule hatte einen Durchmesser von 256 mm entsprechend einer Fläche von 500 cm². Die Eichung der Messfeder ergab pro 1 cm Zusammendrückung einen Lastzuwachs von total 700 kg oder 1,4 kg/cm², bezogen auf die Kontaktfläche zwischen Druckplatte und Eis. Die Ablesung der Federdeformation, die der elastischen Verkürzung des Systems entspricht, erfolgte mittels einer Messuhr und wurde durch zwei symmetrisch angeordnete Messstäbe kontrolliert. Gegen die Stollensohle war die Säule mittels einer Holzbohle abgestützt, deren Fläche ein Vielfaches der oberen Kreisplatte betrug.

2. *Messergebnisse*

Nach Installation der Drucksäule anfangs August 1949 wurde erstmals am 10. August ein Anstieg des Druckes festgestellt. Sein weiterer Verlauf wurde während 6 Monaten durch 1-2 Messungen pro Monat verfolgt und ergab folgende Werte (Tab. 1):

TABELLE 1

Gemessene Eisdrücke auf die Drucksäule

Datum	Ablesung	Δ	P	$\sigma = \frac{P}{F}$	Phase
	cm	cm	kg	kg/cm²	
10. 8.49	19.90	0	0	0	
19. 8.	20.17	0,27	189	0,38	
3. 9.	20.69	0,79	553	1,11	1.
18. 9.	21.10	1,20	840	1,68	
25. 9.	21.46	1,56	1092	2,18	
19.10.	21.67	1,77	1239	2,48	
26.10.	21.74	1,84	1288	2,58	2.
30.11.	22.12	2,22	1554	3,11	
30.12.	22.55	2,65	1855	3,71	
30. 1.50	22.69	2,79	1953	3,91	
17. 3.50	22.70	2,80	1960	3,92	3.

Δ = Verkürzung der Messfeder; P = Federkraft bzw. Eisdruck
σ = spez. Eisdruck; F = Druckfläche (500 cm²).

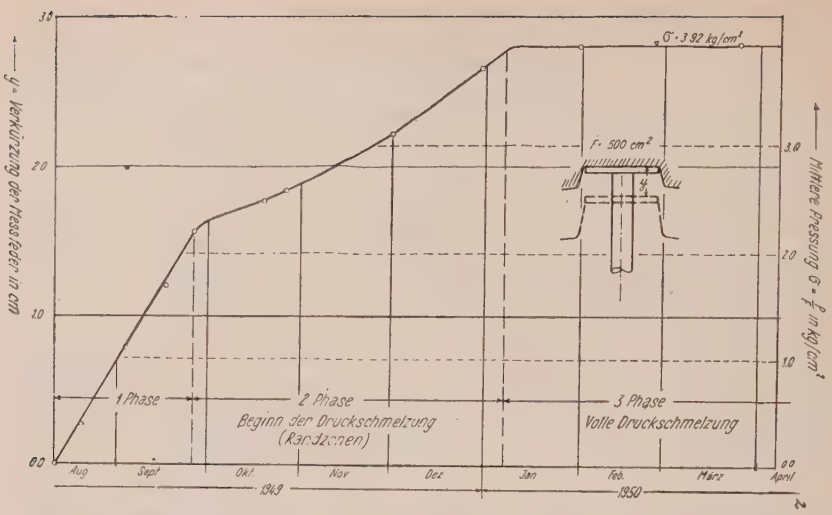


Fig. 2 — Zusammendrückung der Drucksäule und spez. Eisdruck in Funktion der Zeit. (August 1949 - März 1950).

Aus Fig. 2, in welcher die Werte Δ und σ in Funktion der Zeit dargestellt sind geht hervor, dass sich im Verlauf des Eisdruckes 3 Phasen unterscheiden lassen

Die *erste Phase* ist durch einen relativ steilen und gleichmässigen Druckanstieg gekennzeichnet. Die Verkürzungsgeschwindigkeit der Messfeder ist mit 0,028 cm/Tag wesentlich kleiner als die vertikale Kontraktionsgeschwindigkeit des benachbarsen freien Eisstollens (0,1 — 0,15 cm/Tg). Dies lässt auf eine sattere Anpassung der Druckplatte an die rauhe Eisoberfläche durch lokale Regelation und plastische Verformungen schliessen.

Während der *zweiten Phase* stieg der spez. Druck im Laufe von ca. 100 Tagen von ca. 2,3 auf 3,9 kg/cm² an wobei das deutliche Eindringen der Druckplatte in das Eis z.T. unter Druckschmelzung erfolgte. Das Einsetzen der Druckschmelzung bei einer mittleren Pressung, die noch wesentlich unterhalb dem Ueberlagerungsdruck des Eises lag, dürfte durch die Spannungskonzentration an den Rändern der starren Platte bedingt sein.

In der *dritten Phase* erreichte die mittlere spez. Pressung des Eises auf die Druckplatte den Maximalwert von 3,92 kg/cm², der ziemlich genau mit dem Ueberlagerungsdruck an der betreffenden Stelle übereinstimmt. Bei der vorhandenen Eisüberlagerung von 43 m und einem geschätzten mittleren spezifischen Gewicht des Eises von 900 kg/m beträgt der letztere 3,87 kg/cm . Nachdem der gemessene Maximaldruck erreicht war, blieb die Länge der Sonde bzw. der Messfeder konstant.

Die Geschwindigkeit, mit der die Druckplatte unter intensiver Druckschmelzung in das Eis der Stollenkalotte eindrang, entsprach somit der Kontraktionsgeschwindigkeit des Stollens in vertikaler Richtung.

3, *Beurteilung der Ergebnisse*

Wie zu erwarten war, führte der während 5 Monaten beobachtete stetige Anstieg des Eisdruckes schliesslich zu einem Gleichgewichtszustand, bei welchem die elastische Säule sich wie ein starrer Körper verhielt In diesem Zustand war die Relativgeschwindigkeit zwischen der Druckplatte und dem benachbarten Eis praktisch

identisch mit der Kontraktionsgeschwindigkeit des Eisstollens. Der gemessene Endwert des Druckes entsprach dem Ueberlagerungsdruck oder war nur unbedeutend grösser. Dazu sei folgendes bemerkt:

Die grösste Hauptspannung im ungestörten temperierten Gletscher, die im vorliegenden Fall mit dem Ueberlagerungsdruck nahezu identisch war, bestimmt die Temperatur und damit den Schmelzdruck des Eises. Mit dem Ueberschreiten dieses Druckes setzt die Druckschmelzung ein. Analoge Vorgänge spielen sich ohne Zweifel auch an der Gletschersohle zwischen Fels und Eis ab [2] Bei kleinen relativen Geschwindigkeiten dürfte die mittlere spez. Pressung zwischen einem starren Hindernis und dem sich bewegenden Gletschereis weitgehend dem Schmelzdruck als einem unteren Grenzwert entsprechen. Dadurch wird auch der Erosionsdruck, den ein im Eis eingespannter Felsblock auf das Gletscherbett ausübt und von dem die Erosionskraft weitgehend abhängt, limitiert.

Der vorliegende Versuch bedarf als Einzelmessung der sorgfältigen Ueberprüfung mit verbesserter Apparatur. Besonderes Interesse verdient dabei die Frage, wie sich die Intensität der Druckschmelzung und damit die Relativgeschwindigkeit zwischen Druckplatte und Eis mit der Grösse des spez. Druckes ändert.

II. Verformungs- und Geschwindigkeitsmessungen in einer kalten Eiskalotte

Die nachstehend mitgeteilten Ergebnisse bilden eine kleine Ergänzung zu den früher publizierten Beobachtungen und Messungen, die seit 1951 in der Eiskalotte auf dem Jungfraujoch im Gange sind [3]. Dabei handelt es sich einerseits um die kontinuierlich fortschreitende Kontraktion eines quer zur Längsaxe der Kalotte angelegten Kreisstollens und andererseits um den Verformungszustand in einem zentralen Querprofil des ca. 50 m mächtigen Eisschildes.

1. *Verformung eines kreisförmigen Eisstollens.*

Die Lage des 10 m langen Kreisstollens K_2, der einen Teil des im Juli 1954 ausgeführten Querstollens Q_{100} bildet, ist aus Fig. 4 ersichtlich. In der Mitte der kreisförmigen Nische, deren Durchmesser anfänglich rund 2,6 m betrug, wurde die Querschnittsverformung durch die Ausmessung von 8 über den Kreisumfang verteilten Messpunkten verfolgt. Als Messpunkt dienen die Köpfe von ca. 15 cm langen, rund 10 cm ins Eis eingelassenen Schrauben, sodass der merkbare Einfluss der Eisverdunstung nicht in die Messung eingeht.

Die Eistemperatur in der Umgebung des Kreisstollens ist leider nicht genau bekannt und wird für die vorliegende Beobachtungsperiode auf Grund von sporadischen Messungen im Mittel auf ca. — 3° C geschätzt. Die Firnoberfläche liegt rund 25 m über der Axe des Kreisprofils.

Aus Fig. 3, in welcher die gemessenen Verkürzungen verschiedener Stollendurchmesser — ausgedrückt in % ihres Anfangswertes — in Funktion der Zeit dargestellt sind, geht einerseits hervor, dass sich unter den beobachteten Durchmessern der vertikale am raschesten verkürzt. Andererseits zeigt der im allgemeinen lineare Anstieg der Durchmesserverkürzungen mit der Zeit, dass sich die Verengung des Profils mit praktisch konstanter Geschwindigkeit vollzieht. Aus der Differenz der ca. 13 Monate auseinanderliegenden Messungen 7 und 10 ergaben sich z.B. folgende mittlere spez. Verkürzungsgeschwindigkeiten:

Lotrechter Durchmesser $v_v = 3{,}3$ % pro Jahr
Horizontaler » $v_h = 2{,}3$ % pro Jahr

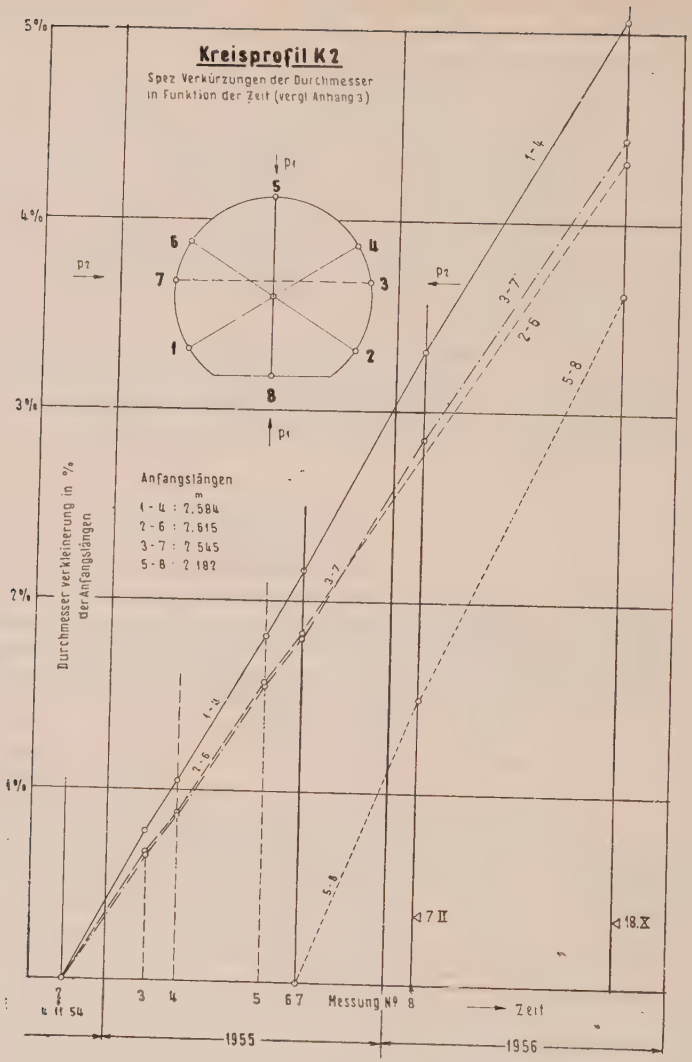


Fig. 3 — Kreisprofil K_2 : Spez. Verkürzungen einzelner Durchmesser in Funktion der Zeit.

Die Richtungen der beiden Hauptspannungen stimmen mit den Richtungen derjenigen Durchmesser überein, welche die grösste bzw. kleinste Verkürzung erleiden. Wie bei den asymmetrischen Randbedingungen der Eiskalotte nicht anders zu erwarten war, weicht die Richtung der in der Messebene gelegenen 1. Hauptspannung von der Lotrechten etwas ab.

Für ein plastisches Medium mit konstanter Zähigkeit lässt sich das Verhältnis der Hauptspannungen $p_2 : p_1$ auf Grund der spez. Verkürzungsgeschwindigkeiten

v der zugehörigen Durchmesser wie folgt berechnen:

$$\zeta = \frac{p_2}{p_1} = \frac{1 + 4\lambda}{4 + \lambda} \; ; \; \lambda = \frac{v_2}{v_1}$$

Da sich jedoch das Eis wesentlich anders verhält als eine Newton'sche Flüssigkeit, kann die Anwendung dieser Gleichung auf den Gletscher nur einen Vergleichswert liefern. Das allgemeine Problem, den Spannungszustand in der Umgebung eines Kreisstollens aus dessen Verformung zu berechnen, müsste auf ähnlichem Wege gelöst werden, wie dies Nye für einen Spezialfall (hydrostatischer Aussendruck) auf der Grundlage des tatsächlichen rheologischen Verhalten des Eises getan hat [4,5].

2. *Horizontale Geschwindigkeiten in einem Querprofil durch die Kalotte (1955/56).*

Nachdem in der Vertikalebene durch den zentral gelegenen Querstollen Q_{100} der Eiskalotte nicht nur die horizontalen Geschwindigkeiten längs der Stollenaxe, sondern auch diejenigen der Kalottenoberfläche in einigen Punkten gemessen worden sind, ergab sich ein erster Einblick in die Verteilung der horizontalen Geschwindigkeitskomponenten innerhalb des betrachteten Querschnittes.

Die entsprechenden Geschwindigkeitsprofile, die mangels genügender Messpunkte nur ein ungefähres Bild der tatsächlichen Verhältnisse vermitteln, sind in Fig. 4 dargestellt. Dabei handelt es sich einerseits um die Verteilung der Horizontalgeschwindigkeiten in der Richtung der Stollenaxe und der Plateauoberfläche und

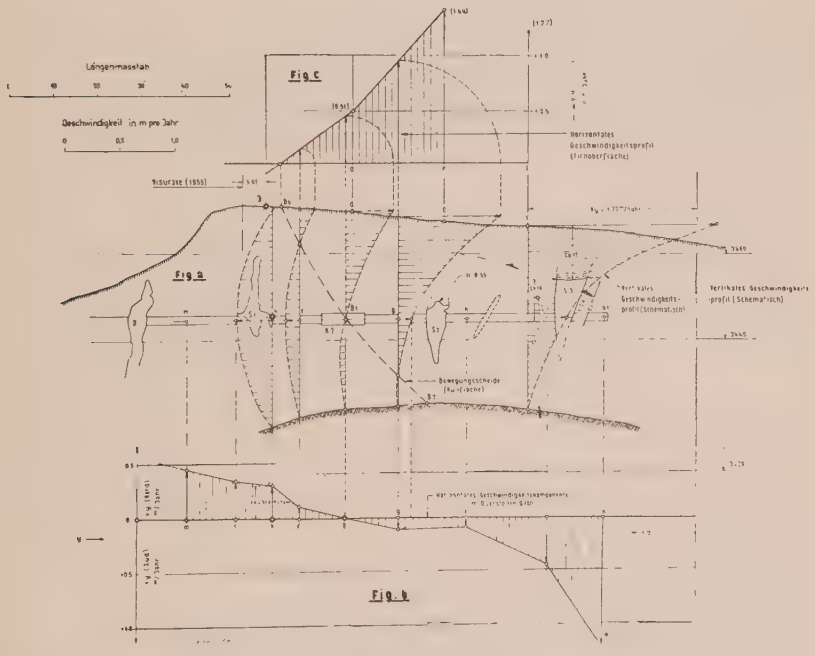


Fig. 4 — Wahrscheinliche Geschwindigkeitsverteilung im zentralen Querschnitt durch die Eiskalotte Jungfraujoch (Schnitt durch den Querstollen Q_{100}).

andererseits um die Erfassung der Geschwindigkeitsverteilung längs einer Lotrechten durch die Kalotte (lotrechte Geschwindigkeitsprofile)..

Die Geschwindigkeitsverteilung längs der Plateauoberfläche wird durch Fig. 4 c veranschaulicht. Es fällt auf, dass die Geschwindigkeitszunahme zwischen dem absoluten 0-Punkt der horizontalen Bewegung (B_o) und dem ca. 57 m nördlich davon gelegenen Punkt F bedeutend grösser ist als zwischen entsprechenden Punkten auf der Höhe des Querstollens. Die spez. Dehnung zwischen den genannten Punkten B_o und F betrug rund 3 % pro Jahr, während sich die senkrecht darunter gelegene Strecke des Querstollens Q_{100} nur um ca. 1 % verlängerte. (pro Jahr)

Das Geschwindigkeitsprofil des Querstollens, das in Fig. 4 b dargestellt ist, zeigt durch seinen unregelmässigen Verlauf, dass die Homogenität des Spannungszustandes durch die Spaltenbildung gestört wird. Bemerkenswert ist der Umstand, dass zwischen den Punkten g und h keine Dehnung beobachtet werden konnte obschon diese Strecke von einer breiten Spalte durchsetzt ist.

Zur Aufzeichnung der Geschwindigkeitsverteilung in der Lotrechten standen zunächst nur die gemessenen oder interpolierten Werte zweier Punkte (Oberfläche und Querstollen) zur Verfügung. Ein dritter Punkt (0-Punkt) ergibt sich unter der Annahme, dass das Eis an der Felsoberfläche festgefroren sei. Diese 3 Punkte erlauben nur eine rein schematische Darstellung einer möglichen Geschwindigkeitsverteilung. Als Tatsache bleibt jedoch die Inversion der Geschwindigkeitsrichtung innerhalb einzelner Profile bestehen, was auf eine eigenartige Form von « extrusion flow » schliessen lässt.

3. *Bewegungs- und Eisscheide.*

Zwischen zwei diametral entgegengesetzten Bewegungsrichtungen kann von einer flächenhaften Bewegungsscheide gesprochen werden. Sie sei definiert als geometrischer Ort aller Punkte, deren Geschwindigkeitsvektoren (Projektion der wahren Geschwindigkeit auf die Bildebene) lotrecht gerichtet ist. Bei asymmetrischen Randbedingungen und räumlicher Strömung bildet die Bewegungsscheide eine krumme Fläche. Diese schneidet die Bildebene in einer Kurve, der in Fig. 4 die Linie B_o —

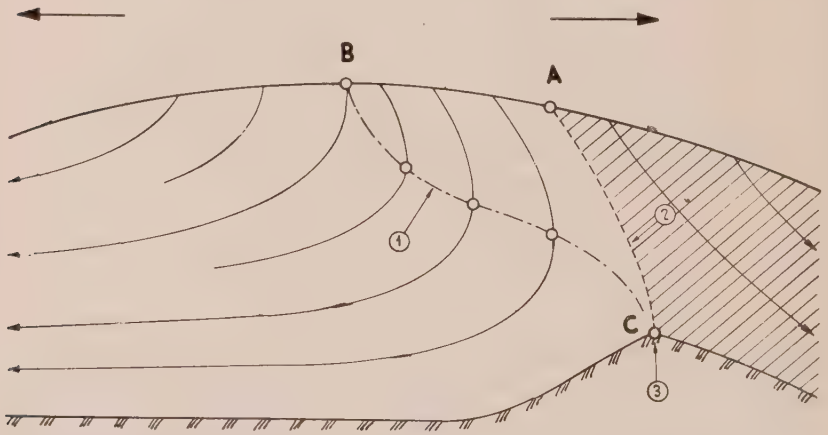


Fig. 5 — Stromlinien in einem asymmetrischen Eisschild (schematisch).
1 Bewegungsscheide (Tangente an die Stromlinien (lotrecht).
2 Eisscheide.
3 Wasserscheide.

B_1 — B_2 entsprechen möge. Von dieser Linie sind nur die Punkte B_0 und B_1 gemessen bzw. reell, während der Punkt B_2 auf der Kulmination (Wasserscheide) der Felsunterlage angenommen wurde.

Wie in Fig. 5 schematisch dargestellt ist, kann man bei asymmetrischen Randbedingungen zwischen einer Bewegungsscheide (1), einer Eisscheide (2) und einer Wasserscheide (3) unterscheiden. Bei der Durchtunnelung eines Eisschildes erleichtert diese Unterscheidung z.B. die Bestimmung einer günstigen Tracéführung mit minimalen horizontalen Verschiebungen. Bei der Beurteilung des Bewegungszustandes des Inlandeises, wo nach den neueren seismischen Untersuchungen die Kulmination B und die Wasserscheide C bis mehrere 100 km auseinanderliegen, stellen sich grundsätzlich ähnliche Probleme [6]. Die vorliegenden Untersuchungen an einer kleinen Eiskalotte mögen dazu beitragen, diese Probleme aufzuwerfen und in einem weiteren Zusammenhang zu sehen.

Die Messungen in der Eiskalotte Jungfraujoch wurden im Auftrag der Generaldirektion der P.T.T. unter der Mithilfe des Ingenieurbureau Rothpletz, Lienhard & Cie. und seines beauftragten Ing. F. Brentani durchgeführt. Die Eisdruckmessungen im Zmuttstollen erfolgten im Rahmen eines von der E O S (S.A. L'Energie de l'Ouest-Suisse) an den Verfasser und an die Versuchsanstalt für Wasserbau und Erdbau an der ETH erteilten Auftrages. Die Konstruktion der Drucksäule stammt von Ing. G. Amberg. Allen beteiligten Instanzen sei für die Erlaubnis zur Veröffentlichung der Versuchsergebnisse bestens gedankt.

Zürich, den 25. Mai 1957.

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THE MECHANICS OF CREVASSE FORMATION (*)

MARK F. MEIER
U. S. Geological Survey
Tacoma, Washington

ABSTRACT

Crevasses are of fundamental importance to an understanding of glacier flow and of fracture patterns in deforming solids. Crevasse patterns on a temperate valley glacier (Saskatchewan Glacier, Canada) were studied as part of a general investigation of the mode of flow of that glacier, and a simple crevasse pattern in the North Greenland Icecap was selected for further work. Preliminary results show that crevasse formation is preceded by a build-up in extension rate. Crevasses then form so as to relieve the extension rate on intercrevasse blocks. In plan view principal strain rate axes are not exactly perpendicular or parallel to the crevasse. Intense deformation in pure shear precedes an extending crevasse. The crevasses studied opened as rigid wedges in a vertical cross section and the jerkiness of opening decreased rapidly with depth.

Some interpretations are suggested. The initiation of crevasses appears to be determined by the curvature of the snow surface and the rate of accumulation or ablation. Relations between spacing and depth of crevasses and prevailing extension rates are given. These relations allow prediction of velocity in areas where direct measurements are impossible. Depths of crevasses can be computed and these results agree well with observation.

Crevasses are a great danger to all who work on glaciers. An understanding of how and why crevasses form would not only decrease their objective danger to glaciologists, but would also be of great fundamental importance to our knowledge of the mechanics of glacier flow and of fracture patterns in deforming solids. Crevasses are usually caused by stresses generated by flow over a nonplanar bed. These stresses can be computed in some very simple configurations. However, existing theories of fracture cannot predict size, shape, spacing, or even the very existence of crevasses in the simplest of known stress situations. To add further complication, the shape, size, and trafficability of a crevasse is also a function of accumulation and ablation around the crevasse. These nontectonic complications will not be considered here.

This paper is essentially a progress report on a continuing investigation of the relation of crevasses to flow in glaciers, and presents data drawn from two separate projects on unlike glaciers (*Meier* and others, 1956, Meier, Rigsby and Sharp 1954. The plan of this investigation is as follows: First, the shapes of forming crevasses in simple patterns and their general glaciological environments are determined. Next, the strain rate fields around the forming crevasses are measured. Finally, attempts are made to relate the strain rate data with the observed crevasse patterns and behaviors. This plan of attack has physical significance because components of the stress tensor are related to strain rate components in an isotropic body deforming by slow plastic flow. Total elastic and plastic strains have some influence on crevasse spacings and patterns, but these strains are not considered in this preliminary analysis.

(*) Publication authorized by the Administrator, Snow Ice and Permafrost Research Establishment, and the Director, U. S. Geological Survey.

Splaying crevasses on Saskatchewan Glacier — Saskatchewan Glacier is a geophysically temperate valley glacier which flows out of the Columbia Icefield in Alberta, Canada. The outlet tongue is 9 km long, averages about 1.4 km wide, is up to 460 m thick, and has a surface slope averaging 3° 23′ (*Meier*, *Rigsby* and *Sharp*, 1954, p. 6-7. A simple pattern of crevasses which occurs on the surface of the glacier 2 to 4 km below the firn limit is shown in Figure 1. These crevasses are longitudinal in

Fig. 1 — Splaying crevasses on a portion of the surface of Saskatchewan Glacier.

trend near the flow centerline and splay out to strike the margin at an angle slightly greater than 45°. This pattern, which is typical of valley glaciers, is termed «splaying». The individual crevasses are V-shaped in cross section, of moderate depth (none deeper than 30 m was observed) and are up to 300 m in length. Because of the moderate depth, rotational strain, and marked surface ablation (2.4 to 5 m per year), few of the individual crevasses are believed to exist for more than 5 years.

The surface strain rate field in this area has been computed using the relations

$$\dot{\varepsilon}_{ij} = \tfrac{1}{2}\left(\frac{\partial V_i}{\partial x_j} + \frac{\partial V_j}{\partial x_i}\right) \qquad (1)$$

where x_i represent a coordinate sytem related to the glacier surface ($i, j = 1, 2, 3$), V_i represent components of the velocity vector parallel to the x_i-axis, and $\dot{\varepsilon}_{ij}$ represent a component of strain rate on a plane perpendicular to the x_i-axis and in the direction of the x_j-axis. These principal axes of strain rate were found to be very nearly parallel or perpendicular to the glacier surface. Therefore, the ice surface is deforming nearly in plane strain. Strain rate components $\dot{\varepsilon}_{ij}$ were derived from measured velocity gradients using relations (1), and were used for the computation of trajectories of principal strain rate in the plane of the ice surface (Fig. 2). It is evident that the

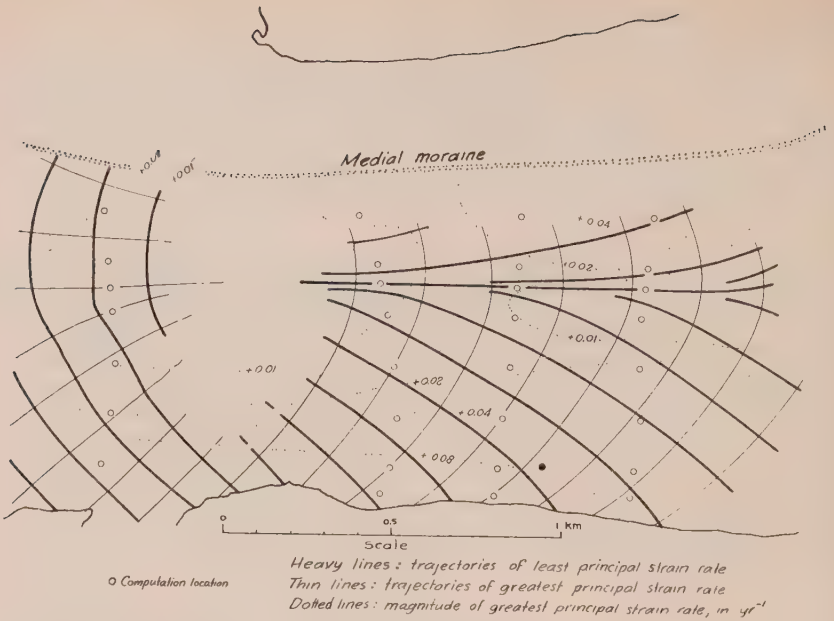

Fig. 2 — Trajectories of principal strain rates on a portion of the surface o Saskatchewan Glacier.

crevasses have formed nearly perpendicular to the direction of maximum principa strain rate (extending). Contours of the magnitude of the maximum principal strain rate show a rough relation to the crevasse distribution, and suggest that crevasses occur only when a principal strain rate approaches + 0.01 yrs^{-1}.

Crevasses in Blue Ice Valley, Greenland — An intensive study of the geometry, mechanics, and thermodynamics of crevassing was begun in 1955 on a small, simple pattern of crevasses in the Icecap of Northwest Greenland (*Meier* and others, 1956). Three parallel crevasses which show a regular progression in length, width, depth, and opening rate were studied, and the third crevasse was instrumented in detail (Fig. 3). These crevasses are in cold firn of density 0.5 to 0.8 $gms \cdot cm^{-3}$. A map of isotherms in a plane across the third crevasse is shown for a typical summer day in Figure 4. These crevasses are similar in appearance to the one described by *Cook* (1956).

Opening rates at depth and profiles of «tectonic width» (Fig. 5) indicate that the third crevasse is opening as a rigid wedge: opening rates are directly proportional to the height above the bottom of the crevasse. The opening at the surface is jerky, but the jerkiness decreases rapidly at shallow depths and then more slowly to greater depths. This behavior appears to be typical of all three crevasses.

The two-dimensional surface strain rate field around these crevasses was measured using relations (1). The velocity gradients used in this analysis were obtained over a short time interval (26 days) so the resulting strain rate data are very approximate. These data, as reported in terms of the orientations and magnitudes of the two principal strain rates in Figure 6, suggest that:

(a) Crevasses do not form until the maximum principal strain rate exceeds + 0.01 yrs^{-1}.

(b) Crevasses tend to relieve the surface strain rate on intercrevasse blocks.

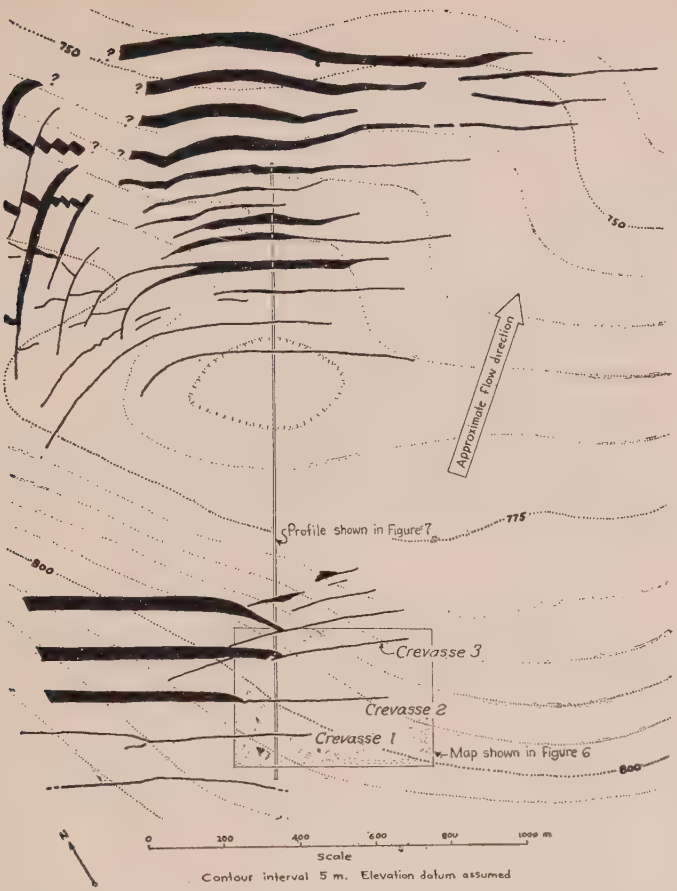

Fig. 3 — Crevasses in the « Blue Ice » area, Northwest Greenland.

(c) Crevasses form at a slight angle to trajectories of maximum principal strain rate.

(d) Shearing displacements across crevasse walls occur.

(e) A flame-shaped area of intense deformation, nearly in pure shear, precedes the actively forming Crevasse 1.

Additional data on surface deformation were obtained from the extension in a year's time of a line of stakes placed in 1954 perpendicular to these crevasses and extending through a noncrevassed area into another area of crevasses. These data were analyzed by assuming that the surface extension adjacent to a crevasse wall was zero, so that all of the extension between two stakes was taken up by crevasse widening. These data are presented in Figure 7 and suggest that :

(f) Extension rate is a function of surface curvature.

(g) An extension build-up precedes crevasse formation, and crevasses do not occur unless the extension rate exceeds about + 0.01 yrs^{-1}.

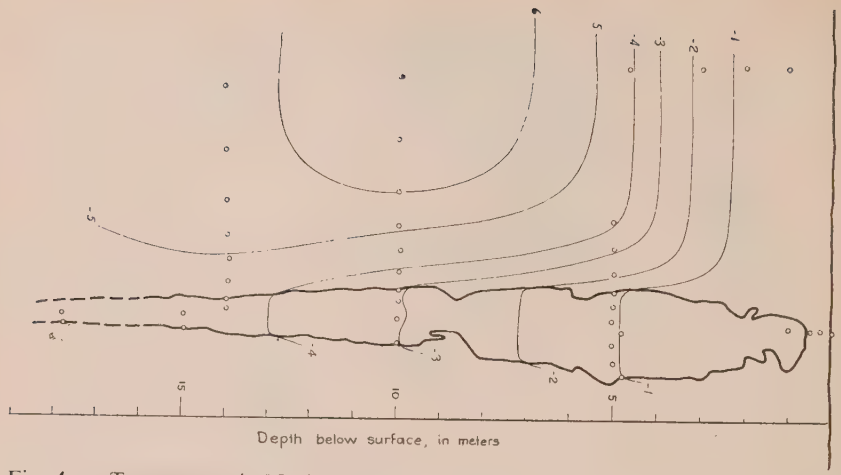

Fig. 4 — Temperature in °C distribution in and adjacent to Crevasse 3, at 6 : 40 AM, August 16, 1955. Data obtained by C. J. Pings, Jr.

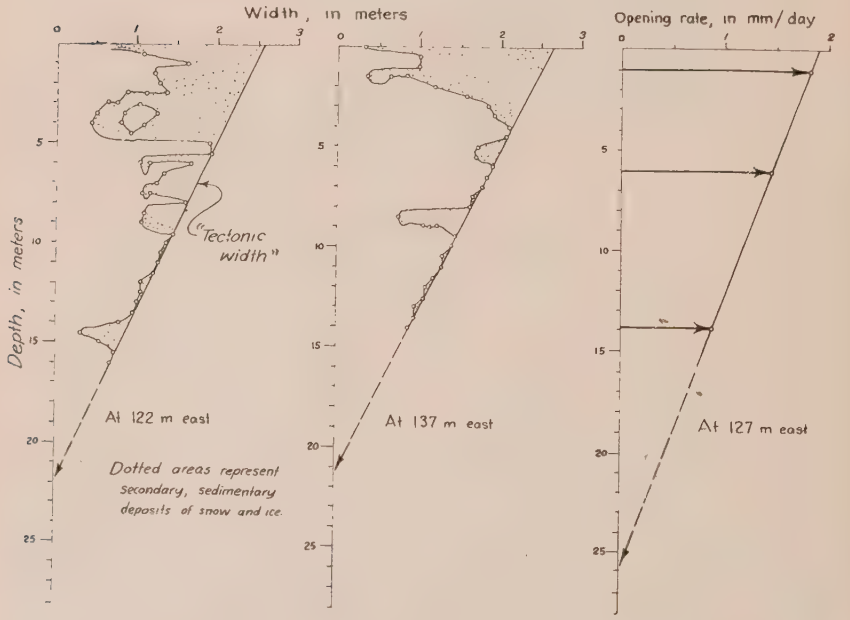
Width, in meters
Opening rate, in mm/day
Depth, in meters
"Tectonic width"
At 122 m east
At 137 m east
At 127 m east
Dotted areas represent secondary, sedimentary deposits of snow and ice.

Fig. 5 — Measured width, «tectonic width», and opening rate as a function of depth in Crevasse 3.

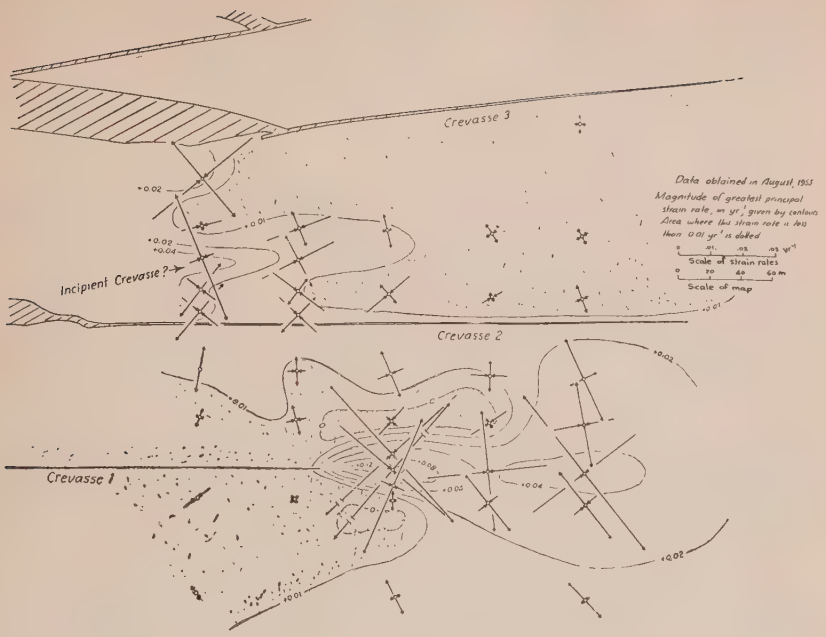

Fig. 6 — Principal strain rates in a portion of the «Blue Ice» area, Greenland

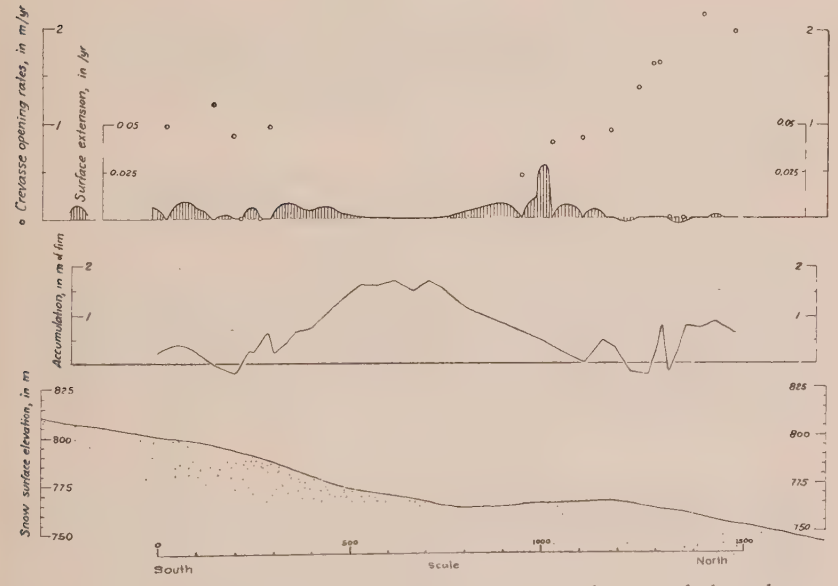

Fig. 7 — Surface extension, crevasse opening rates, and accumulation along a North-South profile in the Blue Ice Area.

(h) Once a crevasse forms, its opening rate either ranges between relatively narrow limits (0.4 to 2 m yrs^{-1}) or it may be completely inactive; no closing crevasses were observed.

(i) Crevasse opening rates are not related to crevasse width.

Interpretation of the Field Data

The condition for the initiation of crevasses — Because of the approximate nature of the field data, it is not yet possible to give an exact criterion for the formation of crevasses. However, the data indicate that crevasses do not form until the maximum principal strain rate on the surface reaches or exceeds about $+ 0.01$ yrs^{-1}. The data suggest that crevasses form at somewhat lower values of strain rate in temperate ice than in cold firn, but the differences are not appreciable. The data from Greenland show that the condition for crevasse formation involves not only normal strain rates but also shearing strain rates. The crevasses on Saskatchewan Glacier show some evidence of shearing at the instant of formation, but this evidence is not conclusive because it cannot be completely separated from the effect of a subsequent rotation. Perhaps crevasses are purely tensile fractures in temperate glacier ice, but involve both tensile and shear stresses in cold firn. Slab avalanches in snow are known to break at an appreciable angle to the principal tensile stress (*Haefeli*, 1948, p. 681), so this fracture condition is not unreasonable for firn.

In the Blue Ice area there appears to be a relation of extension to surface curvature in a direction perpendicular to the crevasses (Fis. 7). This suggests the applicability of *Nye's* (1957, p. 126) formula for longitudinal strain rate $\dot{\varepsilon}$ in a semi-infinite glacier of uniform slope α and constant basal shear stress:

$$\dot{\varepsilon} = \pm \frac{1}{h}\left(\frac{d\phi}{dX} + \frac{\phi}{R}\cot\alpha\right) \qquad (2)$$

where h is the thickness of the glacier, ϕ is the total discharge through a unit width and R is the radius of curvature of the surface. Figure 7 suggests that the accumulation gradient term $\frac{d\phi}{dX}$ is of much less importance than the surface curvature term in the Blue Ice area.

A critical value of principal extending strain rate of $+ 0.01$ yrs^{-1} corresponds to a principal tensile stress of 0.3 or 0.4 bar, according to present knowledge of the flow law of ice. The surface stresses cannot be computed with any pretense of accuracy for either Saskatchewan Glacier or the Blue Ice area.

Crevasse depths and spacings — Crevasses can exist as a stable configuration only to the depth where the longitudinal normal stress in the ice is equal to the stress in the crevasse (atmospheric pressure). *Nye* (1955, p. 513) suggests the following equation for the depth d of crevasses if the flow law of ice is approximated by a power law:

$$d = \frac{1}{\sigma g}\left(\frac{\dot{\varepsilon}}{B}\right)^{1/n} \qquad (3)$$

where σ is the average density, g the acceleration of gravity, $\dot{\varepsilon}$ the longitudinal extension rate, and B and n are empirical constants involved in the flow law. The depths and extension rates of two of the Blue Ice crevasses are known. If *Glen's* (1953, p. 721) values for B and n at different temperatures are used, the following results are obtained:

TABLE 1

Measured and computed depths of crevasses

Crevasse No.	B (yrs^{-1} bars^{-n})	n	Temperature (°C)	d (computed) (m)	d (measured) (m)
3	0.022	3.3	— 1.5	12.2	26.0
	.0086	3.3	— 6.7	15.7	
	.0018	3.3	—12	26.0	
2	.0018	3.3	—12	22.1	22.0

Nye's formula apparently gives correct results if the temperature in the firn at the level of the crevasse bases is —12°C. This value is reasonable because the mean annual temperature at Thule (about 40 km to the east and 800 m lower in elevation) is about —11.4°C (*L. H. Nobles*, personal communication).

Although Nye's approach apparently predicts reasonable depths for these two crevasses, one difficulty is encountered: If a power-law type of flow law is used in this analysis, the opening rates of crevasses should decrease exponentially with depth. This prediction is *not* verified by the measured tectonic width and opening rate data at depth (Fig. 5).

Crevasses form so as to relieve the extension on and in the intercrevasse blocks. In the case of long, straight and parallel crevasses it is reasonable to assume that the depth/spacing ratio is relatively constant when crevasses first form because this ratio is determined by geometrical considerations and the mechanical properties of the ice or firn. In the Blue Ice area, a new crevasse appears to form when the next older one has moved away from the boundary of crevassing a distance equal to about four times its depth. If the crevasse relieves all of the extension on the surface of the intercrevasse block, then the crevasse opening rate $\dot{w}$, the substratum extension rate $\dot{\varepsilon}$, and the spacing s are related as follows:

$$\dot{w} = \dot{\varepsilon}\, s \tag{4}$$

If the crevasses move into an area of greater $\dot{\varepsilon}$, they must increase their opening rates $\dot{w}$ in order to remove extension from the surface of the intercrevasse blocks. By doing this, they increase in depth and thereby increase the depth/spacing ratio. The inverse is not true: a decreasing extension rate can be compensated by a shallowing of certain crevasses to the point where they lose hold on the extending substratum. In this case ,their extension rate would be captured by neighboring, deeper crevasses, and one would observe inactive crevasses alternating with crevasses which were continuing to widen. This is apparently what is happening at the north end of the Blue Ice area (fig. 7).

If the opening rate $\dot{w}$ is relatively constant, a simple relation exists between this and the ice velocity V, crevasse spacing s, and the widths w of the two consecutive crevasses:

$$V = \frac{\dot{w}s}{w_1 - w_2} \tag{5}$$

This formula might be used to measure velocity on glaciers or ice sheets, and can easily be generalized to cover more situations if the variation of opening rate as a function of position is known.

A postulated sequence of crevasse formation and obliteration -- An area is assumed in which the rate of surface extension is sufficient to cause crevasses, surrounded by an area where the extension is not sufficient for crevassing. This area will generally be fixed in space (*Haefeli*, 1948, p. 711) but can expand or contract if the glacier becomes thicker or thinner. The large extension rates within the area would be caused by the factors on the right-hand side of (2) or by large transverse velocity gradients. As ice or firn moves into this area, a crevasse forms and relieves the extension to either side. The crevasse will widen and deepen as it moves so that its base is always at the level where the difference between atmospheric pressure and extending stress equals zero. As this crevasse moves on, extension builds up again at the upper boundary of the area of crevassing until another crevasse forms. If the extension rate is relatively constant within the area, the crevasses will have fixed spacing, depths and opening rates. If the extension rate increases markedly, the crevasses must deepen and widen because the spacing cannot change unless conditions in the third dimension favor the growth of additional fractures.

As a given crevasse approaches an area of decreased extension or is rotated into an unfavorable orientation for absorbing extension, it may lose some of its hold on the extending substratum. Neighboring crevasses which were slightly deeper or more favorably oriented would capture some of this extension. This situation is unstable, and would lead to complete inactivity of the given crevasse. This crevasse would then be eliminated by ablation or accumulation, and as the neighboring crevasses move out of the area of critical extension, the same fate would befall them. This may explain the rarity of closing crevasses.

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REFLEKTIONSSEISMISCHE WIEDERHOLUNGSMESSUNGEN AUF DEM PASTERZEGLETSCHER UND IHRE BEDEUTUNG FÜR DIE FESTSTELLUNG VON GLETSCHER- UND INLANDEISSCHWANKUNGEN

B. BROCKAMP
Münster i. Westf.

Summary

Variations of glaciers can be determined numerical by comparison of cartographical or stereophotogrammetrical maps of different epochs. The seismic or gravimetrical method will be of use, too.

Seismic sounding at the Pasterze — Glacier 1929 and 1955 have demonstrated, that the thickness of the ice on the same profil waned for 62 m during 26 years.

For the identification of the variations of inland-ice it is necessary to make seismic measurements at an area of 2 × 3 km (or so), so that a reliable map of the ice-underground can be obtained. Repeated measurements at the same point will give the variation of thickness, some years later on.

Zusammenfassung

Gletscheränderungen können durch den Vergleich von guten Kartenunterlagen aus verschiedenen Epochen bestimmt werden; aber auch seismische und gravimetrische Messungen können hierzu dienen.

1929 und 1955 auf dem Pasterze-Gletscher ausgeführte reflektionsseismisch Arbeiten zeigen, dass die Eisdicke auf demselben Profil um 62 m abgenommen hat.

Zur Feststellung der Mächtigkeitsänderung des Inlandeises sind flächenhafte reflektionsseismische Messungen notwendig, um eine gute Karte des Untergrundes an einzelnen Punkten des Inlandeises zu gewinnen. Wiederholungen am selben Punkt zu einer späteren Zeit geben die Dickenänderung.

An der genauen Erfassung der in Gletschern und im Inlandeis enthaltenen Eismassen sowie an der Feststellung der Änderungen dieser stillen Reserven im Wasserhaushalt der Erde sind verschiedenene Wissenszweige und, soweit es Gletscher betrifft, auch im steigenden Masse die Technik interessiert und beteiligt. Zur zahlenmässigen Erfassung des weltweiten Gletscherrückganges im letzten Jahrhundert eignet sich, wie R. Finsterwalder ([1]) für mehrere Gletscher der Ostalpen dargestellt hat, die Messung folgender Werte:

1. Lineare Veränderung der untersten Zungenenden in der Fliessrichtung des Eises.

2. Flächenveränderung in verschiedenen Höhenzonen und Änderung der Gesamtfläche des Gletschers.

3. Höhenänderung der Gletscheroberfläche in den verschiedenen Höhenzonen und im Durchschnitt für den ganzen Gletscher.

4. Bestimmung der Höhenlage der Schnee-(Firn) Grenze.

5. Veränderungen im Eishaushalt und im dynamischen Verhalten (Änderung der Eisgeschwindigkeiten) des ganzen Gletschers.

Wünschenswert wäre eine kombinierte Anwendung von 1 — 5, da die einzelnen Gletscher recht unterschiedlich auf die gleiche Klimaänderung reagieren: einerseits

schnelle Verschiebung des Gletscherendes oder seiner Seitenerstreckung. andererseits schnelles Einsinken resp. Anwachsen der Zungenoberfläche, hauptsächlich abhängig vom Profil des Gletscherbettes. 3 ist, da voraussetzungslos, zuverlässig und lässt sich auch verhältnismässig einfach an *Gebirgsgletschern* mit Hilfe genauer stereophotogrammetrischer Aufnahmen aus verschiedenen Zeiten durchführen.

Für grössere *Inlandeisgebiete* gewährt 1 nur beschränkten Einblick in die wirklichen Vorgänge; 2, 3, 4 fallen wegen fehlender Unterlagen aus; 5 hingegen ergibt unmittelbar aus den meteorologischen Elementen brauchbare Werte für den Eishaushalt eines stationären Gletschers im Jahr der Messung: bezeichnet F die Fläche des Firngebietes einschliesslich der Zone des superimposed ice, A die des Ablationsgebietes, z den aus Pegelablesung oder Ausgrabungen und Dichtemessungen abgeleiteten mittleren Zutrag des Gebietes F, a die mittlere Ablation + Eisbergausstoss, so gilt $\frac{F}{A} = \frac{a}{z}$, wobei z, F und A relativ einfach zu bestimmen sind.

Um über längere Zeiten Änderungen in der Höhe der Oberfläche des *Inlandeises* zu erfassen, käme vor allem die Vermessung eines über die ganze Breite resp. Länge des Inlandeises sich erstreckenden Höhenprofils und seine Wiedervermessung nach einigen Jahren in Frage; ein sicher nicht einfaches Unternehmen, sofern Lage- und Höhenangabe nach den klassischen geodätischen Methoden mit genügender Genauigkeit durchgeführt werden sollten (siehe hierzu: H. Lichte ([2])).

Demgegenüber eignen sich zur Feststellung von Mächtigkeitsänderungen sowohl an Inlandeis wie an Gletschern — man könnte fast sagen zu ihrer Überwachung — einige Methoden der angewandten Geophysik, insbesondere gravimetrische und seismische Messungen ([13]); dieses soll an Hand von seismischen Wiederholungsmessungen an einem ostalpinnen Gletscher gezeigt werden.

Auf dem im Glocknergebiet gelegenen Pasterzegletscher wurden 1929 ([3, 4, 5,]) und 1955 seismische Arbeiten ausgeführt. 1929 kam 1 mechanischer Seismograph zum Einsatz, die Registriergeschwindigkeit betrug 81 mm/sec, als Zeitmarkengeber diente eine 50 Hz Stimmgabel; 1955 wurden 15 Geophone + Verstärker eingesetzt. Die Papiergeschwindigkeit betrug 250 mm/sec bei 100 Zeitmarken pro sec. Der instrumentelle Fortschritt bedingt höhere Genauigkeit in den Zeitangaben und wesentlich mehr Material bei dichtliegenden Beobachtungspunkten. Der Vergleich bezieht sich auf den NE Teil der seit vielen Jahrzehnten in regelmässiger Beobachtung stehenden Steinlinie Seelandfelsen-Hoffmannshütte (Seelandlinie), etwa mit den Steinen Nr. 6, 7, 8, 9, 10, 11 sowie 300 m firnwärts und zungenwärts von Stein 9).

Auch wenn bei diesen Arbeiten andere Ziele verfolgt wurden, so lassen sich doch für unsere Fragestellung wichtige Angaben gewinnen.

a) Die Geschwindigkeit der longitudinalen Wellen (P), die für die Tiefenberechnung benötigt wird, wurde 1929 zu 3580 m/sec, 1955 aus wesentlich zahlreicherem Material (1400 Einsätze) zu 3650 m/sec für Längs- und Querrichtung bestimmt. Für den Vergleich wurde eine konstante Geschwindigkeit von 3600 m/sec zugrundegelegt. (s. Weiteres über die Geschwindigkeit seismischer Wellen, am Schluss).

b) Das Längsprofil 1955 verläuft im wesentlichen in derselben Richtung wie das von Standpunkt VII ausgehende leider nur kurze Längsprofil 1929; das gleiche gilt für das von Standpunkt VII 1929 nach SW verlaufende Querprofil.

Auf den Profilen 1929 und 1955 fallen 9 Beobachtungspunkte ± zusammen: die entsprechenden Laufzeiten für die P-und R-Wellen, sowie die aus den R-Laufzeiten berechneten Eismächtigkeiten sind in Tabelle 1 zusammengestellt:

TABELLE 1

Punkt	Δ i.m	t_P i.msec 1929	t_P i.msec 1955	t_R i.msec 1929	t_R i.msec 1955	h i.m 1929	h i.m 1955	Δh i.m
VII zungenwärts	400	113	111	183	155	254	195	60
	500	143	140	202	181	262	201	61
	600	168	167	224	200	268	200	68
VII firnwärts	350	98	96	181	150	250	190	60
	400	110	112	185	155	256	195	61
	500	142	141	202	181	262	201	61
	600	170	167	222	200	265	197	68
VII SW	350	99	98	182	153	252	187	65
	417	116	117	188	166	252	188	64

Die Laufzeiten der P-Wellen 1955 liegen in den in Tabelle I angegebenen 9 Fällen durchweg um 2 msec. früher als die von 1929: sie bedingen so die etwas höhere Geschwindigkeit von 3650 m/sec; es treten aber sowohl positive wie negative Abweichungen auf.

Anders die Reflektionszeiten. Sie sind alle kleiner geworden und zwar in dem untersuchten Entfernungsbereich um 21 — 29 msec. Unterstellen wir für alle R-Einsätze eine Ablesungsgenauigkeit von ± 2 msec, so sind diese festgestellten Änderungen in den R-Zeiten über 10 mal so gross wie die Ablesefehler.

Aus den einzelnen R-Zeiten, die unmittelbar für die « Ueberwachung » des Gletschers benutzt werden, folgt (bei einer Geschwindigkeit von 3600 m/sec) ein in sich übereinstimmender Schwund von 63 m in 26 Jahren (2,4 m pro Jahr)!

Dieser Wert erscheint enorm hoch bei einer Mächtigkeit des Gletschers an dieser Stelle von ~ 260 m (1929). Für die Pasterze ist aber nach V. Paschinger ([6]) gerade das Einsinken der Oberfläche, das sich in der Höhe der frischen Ufermoräne markiert, die auffallende Erscheinung, wohingegen das Zurückweichen des Eisrandes bei der Pasterze viel weniger ausgeprägt ist als bei anderen Gletschern: Wenn sich die Stirn der Pasterze in den 85 Jahren (1898-1942) seit ihrem Maximalstand um 900 m verkürzte, so ist dies, verglichen mit den 1000 m, die der Vernagtferner in 40 Jahren, mit den 740 m, die der Laaser Ferner in 16 Jahren, mit 320 m, die das Maurerkees in 10 Jahren verlor, recht gering zu nennen. Auch die Verschmälerung der Pasterze erreicht mit 160 m unter der Hoffmannshütte nur 1/10 der ursprünglichen Breite.

Diese Ausführungen V. Paschingers lassen grösse Änderungen der Höhenlage der Gletscheroberfläche erwarten. Um so bedauerlicher ist es, dass wegen eines Schlechtwettereinbruchs im Sept. 1955 die für das Profil Hoffmannshütte-Seelandfelsen vorgesehene Höhenmessung unterbleiben musste.

Aber die Nachmessungen am Pasterzekees in den Jahren 1950-1952 von H. Paschinger ([7], [8]) gestatten eine gewisse Nachprüfung unseres Schwundwertes;

a) das Längsprofil von 1955 und das Profil VII 1929 schneiden die Seelandlinie etwa bei Stein 9 des Geschwindigkeitsprofils 1954. Für diese und die benachbarten Steine entnehme ich den Berichten von H. Paschinger folgende Werte:

(Die Entfernungsangaben zählen von Stein 9).

TABELLE 2

	Stein Nr:				
Seelandfelsen	7	8	9	10	11 Hoffmannshütte
Entfernung i.m	197.8	96,9	0	105,8	185.2
Höhenänderung 1949/50 i.m	—3,8	—3,8	—4,7	—5,6	—5,6
Entfernung i.m	206,9	79,0	0	103,7	182,4
Höhenänderung i.m 1950/51	—4,5	—4,8	—2,8	—3,6	—1,7
Entfernung i.m	197,8	97,9	0	107,7	185,3
Höhenänderung 1951/52 i.m	1,9	1,8	2,4	2,2	2,6

Wir sehen, dass in der Tat ganz erhebliche Höhenänderungen in den einzelnen Jahren eingetreten sind.

b) Nach der Glocknerkarte des D.u.Ö. Alpenvereins (1925) schneidet das Längsprofil mit dem Zeltstandpunkt VII (1929) die Seelandlinie bei 2397 m, am Lagepunkt des Steines Nr. 9 1954.

Für Stein 9 1952 gibt H. Paschinger einen Höhenwert von 2345 m an. Nehmen wir für die Zeitspanne von 1952-55 einen jährlichen Schwund von 2 m an, so würde für 1955 eine Höhe von 2339 m folgen, also gegenüber 1955 eine Abnahme der Höhe von 58 m; Stein 7 und 8, für die eine mittlere jährliche Abnahme von 3 m für 1943-1952 sich ergeben und unter denen die Reflektionspunkte der beiden Querprofile liegen, würden eine noch bessere Uebereinstimmung mit den seismischen Werten zeigen.

c) In der eingangs erwähnten Arbeit (l.S.211) gibt R. Finsterwalder auch die jährliche Höhenänderung als Funktion der Höhe (1920-1950) für 8 Gletscher der Ostalpen; für eine Höhe von 2350 m folgt eine mittlere Abnahme von 1,80 m; Sülzenauferner und Schwarzensteinkees zeigen in 2350 m Höhe eine Änderung von etwa 2,50 m; sie ist für diese Gletscher von gleicher Grösse wie sie aus den seismischen Messungen auf der Pasterze folgt.

Auch wenn bei der Anlage der seismischen Arbeiten 1929 ein Vergleich mit späteren Untersuchungen nicht ins Auge gefasst war, so konnte doch trotz unterschiedlichen Beobachtungsmaterials die Brauchbarkeit der seismischen Methode zur Feststellung der Änderung der Gletschermächtigkeit unter Beweis gestellt werden; der abgeleitete enorme Schwund der Eisdicke von 63 m in 26 Jahren, der rund 1/4 der Eisdicke von 1929 ausmacht, steht mit der Höhenänderung, wie sie aus mir zugänglichen Kartenunterlagen und Vermessungsdaten folgt, in guter Uebereinstimmung.

Seismische resp. gravimetrische Wiederholungsmessungen vermitteln über die rein zahlenmässige Feststellung der Höhenänderung von Gletscher-resp. Inlandeisoberfläche hinaus, auch Angaben, um wieviel Prozent sich die Eisdicke geändert hat. Liegen für mehrere Punkte des Akkumulations- und des Ablationsgebietes derartige Wiederholungsmessungen vor, so lassen sie weitere Schlüsse auf die Eisdynamik zu.

Zur Erfassung dieser Mächtigkeitsschwankungen des Inlandeises wird man zweckmässigerweise an einzelnen Punkten des Inlandeises, deren geographische Breite und Länge bestimmt wurde, mit möglichst steilen Reflektionen flächenhaft arbeiten, um eine Karte des Untergrundes zu erhalten.

Bestens werden solche Stellen gewählt, deren Oberflächenform stärkere Unebenheiten des Felsuntergrundes erwarten lassen. Es werden so Fixpunkte unter dem Eis gewonnen, auf die bei den Wiederholungsmessungen, bei der Bestimmung der

Fliessgeschwindigkeit des Eises usw. alle weiteren Beobachtungen bezogen werden können.

Bereits auf der Deutschen Grönlandexpedition Alfred Wegener ([10]) wurde bei km 120 (71° 12,6′N, 47° 45′W) seismisch flächenhaft beobachtet. Wenn auch für diese 2 × 3 km grosse Testfläche nur 11 mit mechanischen Seismographen gewonnene Reflektionen zur Konstruktion der Untergrundkarte zur Verfügung standen, so ist doch für dieses Gebiet eine Wiederholungsmessung auf der Internationalen Grönlandexpedition (EGIG) beschlossen worden. Wiederholungsmessungen sind auch an einigen seismischen und gravimetrischen Arbeitsplätzen der französischen Grönlandexpedition P.E. Victor vorgesehen.

Es sei an dieser Stelle noch kurz auf die Geschwindigkeitsbestimmungen der longitudinalen Wellen auf der Pasterze eingegangen. Die Geschwindigkeit ergab sich 1929 zu 3,58 km/sec. Die Auswertung von 1400 P-Einsätzen 1955 ergab, dass die Geschwindigkeit in Längs-und Querprofil bis 7-8 m Tiefe 3,48 km/sec, darunter 3,67 km/sec beträgt.

Diese Unstetigkeit in 7 m Tiefe wurde durch Messungen mit 2 Echolot-Apparaturen (30 KHZ,λ = 12 cm; 15 KHZλ = 25 cm), bestätigt. Diese Lotungen, die an mehreren Punkten des Seelandprofils und an verschiedenen Orten unterhalb dieser Linie ausgeführt wurden, ergaben ausser mehreren anderen reflektierenden Flächen einen eindeutigen Reflektionshorizont in 7-8 m Tiefe, also in derselben Tiefe, in der die seismischen Longitudinalwellen (λ = 45 m) ihre Geschwindigkeit ändern. Die kleinere Geschwindigkeit in der oberen Schicht dürfte durch höhere Porosität bedingt sein. Aus den Geschwindigkeiten würde ein Schmelzwassergehalt von etwa 5 % in den oberen 7-8 m der Pasterze sich ergeben, falls der gesante Porenraum nur von Wasser erfüllt wäre.

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CURIOUS OPEN WATER FEATURE IN THE ICE AT THE HEAD OF CAMBRIDGE FIORD

MOIRA DUNBAR (*)

ABSTRACT

In April 1951 the RCAF photographed a pool in the fast ice of Cambridge Fiord which formed such a perfect circle, and had such clean edges, that it was at first mistaken for a spot on the negative. The area was photographed at intervals for the next two years, and each year the pool reappeared in an otherwise continuous cover of fast ice. It varied from about 150 to 300 feet in diameter and appeared, in one year at last, quite late in the season, a photograph in February showing no sign of it. An attempt will be made to analyse the evidence in the photographs, to correlate it with the available meteorological and hydrological data, and to suggest possible causes for the phenomenon.

On 30 April 1951 an R.C.A.F. aircraft on a training flight photographed an open pool at the head of Cambridge Fiord in northeast Baffin Island. This pool or hole in the otherwise complete fast ice cover of the fiord was so perfectly circular that it was taken for a blot on the negative until members of the crew insisted that they had seen it with their own eyes. (Fig. 1) It was about 312 feet across and close to shore at a place where the coast made almost a right angle, the centre of the pool being 600 feet from the shore on each side of the angle.

In the two following years it was possible to get further photographs of the area during the late winter. In 1952 it was photographed six times between March 5 and May 29, and in 1953 three times between the middle of February and the end of April. In each year the pool reappeared in exactly the same place. It was always circular, though never quite so geometrically perfect as in the first photograph, but its diameter varied somewhat from year to year. Thus on 30 April 1951 it was about 312 feet, on the same date in 1952 about 208 feet, and on 29 April 1953 only 156 feet. In 1952, the only year in which it was possible to trace its development, it grew from 112 feet on March 5 to 229 feet on May 29, and a narrow crack gradually spread shorewards from it. (Fig. 2) This crack was about 200 feet long on April 2 and had connected with the shore by April 17, with a small opening running south-eastwards from it along the shoreline. A very interesting point was noted in 1953, when photographs taken on February 17 and March 15 revealed an unbroken ice-cover with no sign of a pool, which however appeared between then and April 29, the date of the next observation. Thus it apparently does not remain open all through the winter, but opens up late in the season, the date varying in the years of observation from before March 5 (the earliest date it has been seen) to later than March 15.

In 1957, after an interval of three years, the pool was again photographed. On May 20 it was about 218 feet across, with a crack joining it to the shore as in 1952, but running to the western instead of the southern shore. (Fig. 3) As in 1952 there was also a narrow crack at the shoreline. In at least one of the three intervening years, 1954-6, the pool was observed visually though not photographed. There is therefore little doubt it is a regular annual phenomenon.

The question of what causes this opening in the ice is an intriguing one to which there seems to be no easy or obvious solution. In the absence of any field observations it is not possible to attempt to find an answer, merely to present the problem and

(*) Miss Moira Dunbar, Defence Research Board, Ottawa.

suggest a few possible lines of approach. Open patches as such are not uncommon; they occur for instance in Frobisher Bay and Cumberland Sound, where they are apparently due to water turbulence, and in many other parts of the Arctic where strong currents keep ice from forming. There is no other known case in the Canadian Arctic, however, of an open pool at the head of a long narrow fiord; in the many such fiords in eastern Baffin Island, Ellesmere Island, and Axel Heiberg Island the ice cover is normally complete and often very smooth, suggesting that it forms in

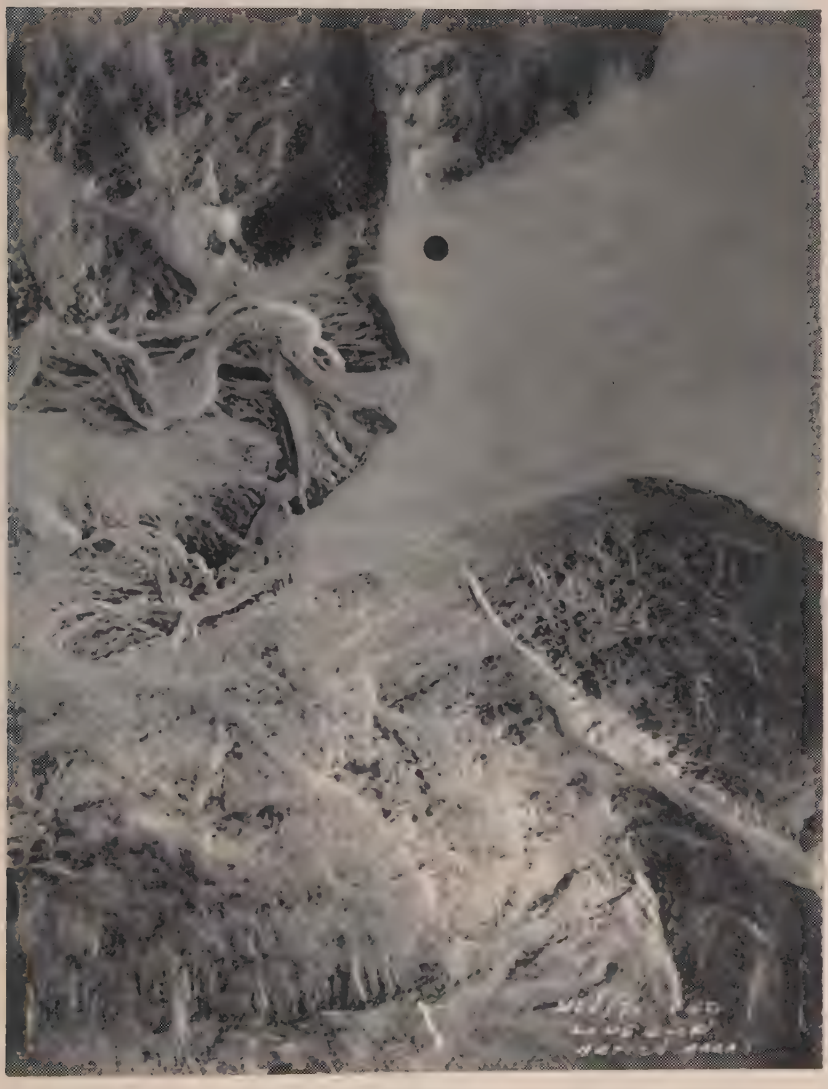

Fig. 1 — The pool as first photographed, 30 April 1951. Vertical photograph taken from 10.000 feet.

quiet water. Furthermore, in places where open pools occur, there are normally several rather than just one, and they are usually larger and quite irregular in outline.

Fig. 2 — 2 April 1952, taken from 6.000 feet. Note the creek beginning to form which later extended to the shoreline.

Another quite abnormal feature of this pool is its appearance late in the season. Pools caused by tidal currents or other forms of water turbulence owe their existence to the fact that either mixing or sheer speed of water movement prevents ice from forming; thus they remain open from the time of freeze-up, and if they freeze over at all it is late in the winter, when accumulation of frost has chilled all the available water to freezing point.

For these reasons it seems unlikely that the feature is a purely oceanographic one. Its proximity to shore, too, leads one to surmise that it may be more closely connected with the physiography and hydrology of the area. Let us therefore consider what little information is available on this subject.

The area is part of the uplifted eastern edge of the Canadian Shield, and appears to be entirely Precambrian, consisting of gneissic rocks. (Fig. 4) The fiord is one of the longest of eastern Baffin Island and penetrates through the mountainous coastal

Fig. 3 — Oblique view looking southwest. 20 May 1957, altitude 5.970 feet.

belt, which rises in this area to about 5000-6000 feet, to the plateau of central Baffin Island, about 2000 feet. Minor faults are common and the uplift of the area has left a series of terraces so well developed that J.M. Wordie, who visited the fiord by ship in 1937, originally called it Glen Roy Fiord, after Glen Roy in Scotland, the classic example of terrace development. The highest terraces observed by Wordie he judged to be at about 1500 feet, and he considered these to have been formed in an ice-dammed lake. No geological results of this expedition were published, and the only man to visit this area since then was G.W. Rowley, who made a winter journey across Baffin Island by way of the Rowley River in 1939, emerging into Keel Bay. He did not unfortunately pass close to the site of the pool and made no geological observations.

Climatically the area is presumably influenced by the proximity of Baffin Bay, but the maritime effect will be modified by its distance from the coast and the mountains which intervene. Of the meteorological stations available, Clyde and Arctic Bay, neither is strictly comparable; the climate is likely to be somewhat between the two, with mean winter temperatures of about —15 to —25°F and summer temperatures about 40-45°F; annual precipitation about 6 inches; and mean cloudiness about 60%. In late winter the weather, as in most of the Canadian Arctic, should be mainly cold and clear. A comparison of meteorological records for Arctic Bay and Clyde for the years in question shows no apparent correlation with the dates of appearance or size of the pool.

The theoretical period of continuous sunshine is from May 12 to August 8, and the sun is continuously below the horizon from November 21 to January 23.

The mountainous terrain will greatly modify these dates however; the spring reappearance of the sun, for instance, will be retarded several days. The pool thus makes its appearance about 4 to 6 weeks after the sun rises above the horizon, but it is hard to see any connection here either. It is much too early in the season for surface melting, and no signs of any appear in any of the photographs. The ice at this period is indeed still growing and has not reached its maximum thickness. Ice measurements made by Bernier at Pond Inlet in 1906-7 and at Arctic Bay in 1910-11 show a maximum thickness of 56-61 inches and a decrease beginning about the end of May or beginning of June, and it is likely that conditions in Cambridge Fiord are somewhat similar.

It would appear that the explanation must involve the existence for one reason or another of a narrow column of water welling up under the ice. This could, of course, be a purely oceanographic phenomenon; nothing is known of the depth of water or other oceanographic factors. For reasons already stated, however, a land-water origin seems more probable. The possibilities seem to be:

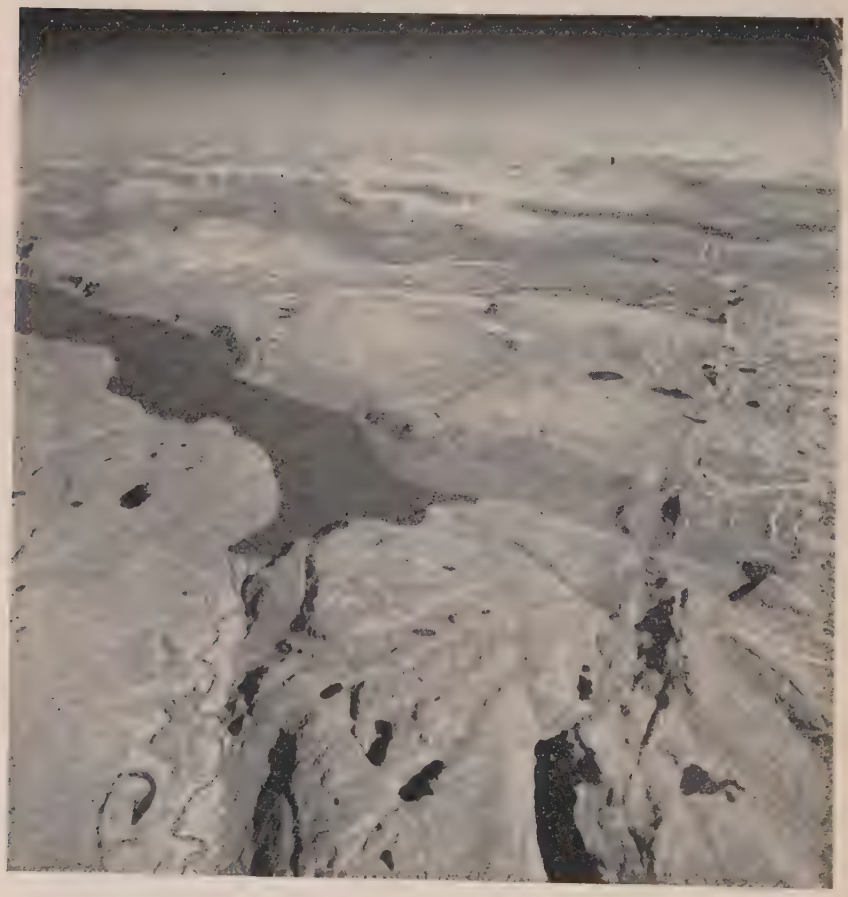

Fig. 4 — Cambridge Fiord looking east, showing the general terrain. Keel Bay is the left bay of the three indentations at the head of the fiord, and the site of the pool is in the centre one. Altitude 20,000 feet.

(1) Surface or sub-surface run-off under the ice
(2) Underground water.

The main drainage into the fiord is by the river flowing into Keel Bay; the stream that enters the fiord near the pool is a short one originating in a lake only about five miles inland. There is no sign of open reaches in any of the winter photographs of this stream, and the wide mouth of the stream freezes over very smoothly, with the surface generally swept clear of snow; but photographs taken in September show what appears to be a fair flow of water about 30 feet wide, and it may well be that the flow continues all winter under the ice. The hole is not, however, opposite the mouth of the river but very much to one side, and there is no obvious reason to connect the two. Furthermore surface run-off would be unlikely to result in a perfectly round pool.

There remains underground water in the form of springs, either hot or cold. The area is one of hard impervious rock, but the many faults suggest that springs from below the permafrost are by no means improbable. And a spring could provide the kind of narrow, even column of water necessary to produce such a pool. The late winter opening however remains puzzling. It might be expected that any spring, hot or cold, would keep the pool open all through the winter. The late opening presupposes not merely that the flow is insufficient to keep the area from freezing over in the fall, but that it then becomes strong enough to melt through the ice that has already formed. This ice may be relatively thin, forming a weak point in the ice cover, but some change in the force of the flow would be necessary to cause it to open up. A possible explanation might be that freezing of the active layer at this time of year reaches a stage that in some way affects the hydrostatic pressure and increases the flow of the spring.

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SCOPE, STATE AND DEVELOPMENT OF PRECISE GLACIER SURVEYS ON THE EARTH

R. FINSTERWALDER
Munich

ABSTRACT

Precise glacier surveys are of special value, as they enable us not only to fix position, height and surface of glaciers but also in the case of repeated measurement to know the variations of glaciers and the change of climate. This was shown by the author in the case of Eastern Alps.

The accuracy necessary for the heightmeasurement of the whole glacier surface s about 1-2 ft.; several points have to be fixed with higher accuracy. The surveys should be done at the time of the yearly minimum of snow-cover above the snow line.

As method for the survey photogrammetry is convenient, in high mountains especially the terrestrial method, in the arctic or antarctic regions with their flat icecaps aerialphotogrametry is more useful; but the farther distance is from the ice borders the greater are the difficulties for precise survey methods.

In the following a short review is attempted of such glaciers which are precisely surveyed especially of those where measurements have been repeated. But it has to be said that the review is rather preliminary and uncomplete; a complete review on an international scale should be the task of the Commission on Snow and Ice.

1. Repeatedly surveyed:
Western Alps: Rhone-, Unteraar-, Aletschglacier.
Eastern Alps : Eight typical glaciers in Zillertal-, Stubai- and Oetztal Alps.
Himalaya: Baltoro- and Rakhiotglacier.
Greenland: Equip Sermia on the western coast.
2. Once surveyed:
Western Alps: several.
Eastern Alps: 14 typical glaciers in the Geophysical year 1957. Aeral Surveys 1952 of most Austrian glaciers.
Norway : Glaciers of Jostedalsbree, Spitzbergen.
Himalaya · Baltoro-region, Nanga-Parbat-region, Kangchendzönga-region, Hunza-Karakorum.
Pamir: Fedtschenko region.
Northern America : Cascade Range, Mt. Rainier.
Southern America: Cordillera Blanca, Cordillera of Huay Huash.
Africa: Ruwenzori-mountains, Mt. Kenya.

Results: In the Alps. 0,40 m/year medium loss of height of glacier surface between 1920-1950.
Rise of snow-line: 60-90 m from 1920-1950.
In the last three years : More accumulation in the firn regions 1-3 m in the whole.
Himalaya — Nanga Parbat, Rakhiot-glacier: Retreat since 1924, but now again growing (greater ice-velocity).
Future Development: Owing to difficulties caused by wet and cloudy summers choose a little test-glacier which is surveyed every year; the other glaciers to be surveyed in the middle each 5th year.

Precise glacier surveys are of special value as they enable us not only to fix position, height and surface of glaciers but also in the case of repeated measurements to know the variations of glaciers and the change of climate. Additional measurements as for instance of ice velocity and of the depth of the glacier, the latter by seismic soundings, give furtheron valuable data.

The author has explained in an article « Die zahlenmässige Erfassung des Gletscherrückgangs an Ostalpengletschern » in which manner one gets from the precise and complete measuring the surface of whole glaciers, the variation of glaciers and

climate, for instance the raising of the snowline. In this paper it is discussed how we can attempt to perform such work over the whole earth, what results have been hereby already reached and what results we can aspire and how we can prepare them.

As «precise» glacier measurements we understand in the following such surveys by aid of which it is possible to fix the glacier-surfaces. This is done during the yearly snow-minimum above the snow-line with an accuracy of height of 1-2 ft over the whole surface. Several points ice- or snowgauges should be fixed still more precisely. Higher accuracy is not necessary for fixing of the surface as one single snowfall can cause alterations of the surface height in this amount. Out of the same reason one cannot obtain generally the yearly snow-minimum with higher accuracy. It would also be rather difficult and very expensive to make measurements more precisely over extended areas especially in big firn basins. On the other hand the accuracy of 1-2 ft is necessary in order to fix the relative small alterations of height in the firn regions; the amount of them is only a little fraction of those in the tongue areas but it is of special glaciological and climatic interest to know these small alterations in the firn region..

In order to survey glacier-surfaces, as a measuring method above all photogrammetry comes into question, and that as well terrestrial as aerial photogrammetry. The first is of advantage especially in *high mountains*, as on uses fixed standpoints on solid terrain near the glacier. Triangulation for determining the coordinates can be easily connected with the photogrammetric work. In the case of later repetition of the photogrammetric survey it is of decisive value, that the standpoints can be relocated mostly without difficulty by aid of the former photographs. In the case of air-photogrammetry fixing and identification of the necessary control-points is very difficult. The determination of the control-points requires a flow of work entirely separated from the aerial survey. Signalization of points above the snowline and identification of such points is that difficult, because marking and survey-flight are performed at different moments, and as in the meantime the limits of snow, which are often near the points, change very noticeably. — The largely extended firn-areas are often rather white and do not show any crevasses; they give very poor or in several cases no possibility for setting the *seen* floating mark stereoscopically on the ground. As a rule terrestrial photogrammetry is also in this respect of advantage, as it is more versatile for example it allows to wait for better light-conditions.

In the *flat glacier regions* of the polar-icecaps as a rule aerial photogrammetry is the method to be considered if it is possible at all to overcome geodetically the huge ice- and firn areas. With increasing distances from the ice-borders the possibilities of precise glacier measurements deteriorate, as also meteorological conditions are not common, and as the atmospheric refraction is exceptionally great and irregular. Therefore normal geodetic methods for fixing points and controlpoints fail or are rendered extremely more difficult than usual.

Measurements of ice-velocity are important, as growing and decreasing of the glaciers are of influence also on the ice velocities. Measurements of ice velocity can be performed photogrammetrically on glacier tongues if they are quickly moving, on the other hand geodetic intersection-methods can be taken into consideration using points on solid terrain near the ice border as standpoints. The measured points on the ice are fixed by poles drilled into the ice or by other signals. The poles can be also prepared for measuring the ablation. The possibility for accurate velocity-measurement deteriorates with increasing distance of the solid ice-border. In addition it has to be considered to determine only the relative velocities especially in order to get the divergence or convergence of the streamlines. In the case of velocity measurements over long time-intervals, perhaps over one or two years, the distances covered by the ice are so large that eventually most precise astronomical measurements can be performed in order to fix the starting and the end point of the movement.

In the following the attempt is made to give a summary of those glaciers which are already measured with the accuracy mentioned before. Of special value are those glaciers the measurements of which are repeated, systematically worked out and put together.

But it is emphasized that our summary is by no means complete. On the contrary it is suggested that the Commission on Snow and Ice initiates the completion of the summary on an international scale.

1. *Repeatedly surveyed glaciers*

Western Alps: Rhone-Unteraar-, Aletschglacier
Basic publication « Vermessungen am Rhonegletscher 1874-1915 mit zahlreichen Karten », Basel 1916.

Eastern Alps: Eight typical gladiers of different size. Several repetitions between 1880 and 1955. R. Finsterwalder « Die zahlenmässige Erfassung des Gletscherrückgangs an Ostalpengletschern ». Zeitschr.f. Gletscherkunde Bd. II, Heft 2/1953, S. 189-239 with contour plans of the glaciers and their states.

Himalaya: Rakhiot-glacier in the Nanga Parbat group. Surveyed in 1934 and 1954. W. Pillewizer « Der Rakhiot-Gletscher am Nanga Parbat 1954 » Zeitschr. f. Gletscherkunde Bd. III/2 1956, S. 180-194.

Greenland: Tongue of Equip Sermia in Western Greenland. Several times surveyed at last by Expeditions Polaires Françaises.

Spitzbergen: Glacier Goes, Gänsegletscher. Surveyed 1899 and 1938. W. Pillewizer. Petermanns Geogr. Mitt., Erg. Heft Nr. 238, S. 14.

2. *Once surveyed glaciers.*

Existing official maps. The official maps of Germany, Austria, Switzerland, Italy and France are surveyed in large scales 1:25 000 or 1:50 000. Notwithstanding they are in general not convenient for accurate glacier investigations. In surveying and compilation of these maps it was not possible to take care of glaciological aspects and the special conditions of the glaciers. Though for mere topographical purposes they are sufficient, the accuracy of 0,50 m is generally not reached for the heights of the surfaces in the firn areas. Also the limits between snow and solid terrain are very often not so well and reliably worked out as it is necessary for glaciological investigations. Very often the surveys are not done at the time of the yearly snow-minimum, therefore the maps do not correspond to the actual yearly state of the glacier. Often the date of the survey is not indicated on the map.

Strips of aerial photographs have been made 1953 by the Austrian Bundesvermessungsamt over the entire Austrian Alps nearly at the time of snow minimum. They however have not yet been evaluated for glaciological purposes in particular.

It would be desirable to find out those glaciers, which are really precisely measured by official topographic surveys; if there would exist a great number of such glaciers one had to chose typical glaciers for further glaciological measurements. At the end of this article something is said about this choice.

In the following list mainly special photogrammetric surveys of high accuracy fit for glaciological investigations are enumerated:

Norway: Nigardsbre, Toerbre and Bergsäterbre in Jostedal surveyed 1937 in 1:25 000. Published Allgemeine Vermessungsnachrichten 1951. « Erd- und Luftphotogrammetrie im Gebirge. Zur Karte des oberen Jostedal in Südnorwegen ».

Spitzbergen: Some typical glaciers in the Hornsund-regions. Map 1:25 000.

W. Pillewizer: « Die kartographischen und gletscherkundlichen Ergebnisse der deutschen Spitzbergen-Expedition 1938 ». Peterm Geograph. Mitt. Ergänzungsband 238, 1939.

Himalaya:

Karakorum-Baltoro-Glacier. Work and maps of the Italian Research-Expeditions of the Duke of Abruzzi and Prof. Desio. Data. of publications will be supplied later.

Himalaya-Nanga Parbat: « Die Gletscher des Nanga Parbat » 1934, Map 1:50 000. Zeitschr. f. Gletscherkunde Bd. XXV 1937, S.57-108.

Zemu-Gletscher: Map 1:33 333 by K Wien, published in the book « Um den Kantsch » by Paul Bauer 1933.

Pamir: Fedtschenko-glacier and the others glacier of the Seltau-group and Transalai-chain. Map 1:50 000. Wissenschaftl.Ergebnisse der Alai-Pamir Expedition 1928, Bd.I u. II D. Reimer, Berlin 1932. Ferner K. Wien « Die Gletschergebiete der Pamire und Westturkestans ». Zeitschr.f.Gletscherkunde XXIII, 1955, S.36-56.

In work, under compilation. Map of *Hunza-Karakorum* especially of the *Batura-glacier* surveyed 1954 by Pillewizer and Heckler+ 1:50.000 Map of Everest-group 1:25 000 surveyed 1955 by E. Schneider.

North-America: Nisqually-Glacier on Mt. Rainier surveyed 1952 , Map 1:25.000 W. Hofmann. « Zur Karte des Nisqually-Gletschers » Z.f.Vermessungswesen 1955, S. 198 — Repetition of the survey and extension over all glaciers of Mt. Rainier 1956 by W. Hofmann.

South-America:

Cordillera Blanca: Huascaran group surveyed 1953 by W. Hofmann.. Map 1:25 000 under compilation.

Cordillera of Huayhuash. Map 1:50 000, surveyed 1936 H. Kinzl, F. Ebster, u.E.Schneider. « Die Karte der Cordillera von Huayhuash ». Zeitschr. d.Ges. f. Erdkunde, Berlin 1942, S. 1-35.

Africa:

Ruwenzori — mountains. Map surveyed in 1:25 000 in 1938. A. Stumpp. « Kartierungsarbeiten im Ruwenzori-Gebirgte ». Bildmessung und Luftbildwesen 1952, S. 142, Allg. Vermessungsnachrichten 1952/6.

Mt. Kenya: Lewis-glacier surveyed 1924, mapped in 1:13 333. C. Troll: « Der Gipfelgletscher des Mt. Kenya. Glaciers and Climate ». Geografiska Annaler 1949. S 257-274.

Summarizing it is to say, that on some essential places of the earth the base has been established for obtaining the variations of glaciers and the climate. But it is necessary to repeat the surveys. In principle these repetitions can be performed with less work and expenses than the primary survey, as the basic trigonometric measurement for fixing stand- and controlpoints have not to be repeated.

Results. On some places we have already now the possibility to characterize glacier and climate variations and this on the base of repeated measurements.

1.) *Alps*: Since 1920 the glaciers retreat again. The medium loss of height of the surface of a whole glacier is 0,40 m/year. During the last three years 1953-1956 the retreat has come to a standstill. Three cool and wet summers have caused even an increase of the firn-surface. It amounts on the firn-line to about 1,20 m. In higher regions the amount is about 5 m, in a height of 500 m above the firn-line. — The tongues of long glaciers are still retreating; on short active glaciers a little advancing

has already influenced the tongue. It is to suppose that the glacier-retreat has been stopped but temporarily.

Between 1920-1950 the snow-line raised about 60-90 m.

2.) Himalaya — *Nanga Parbat.* Since 1934 a relatively small retreat of the tongue is noticeable. Its end has retreated about 500 m. In the moment one can conclude from measurements of the ice velocity to an advance. The velocity is about 20% larger than 1934. As the results have been obtained by one glacier only they are somewhat uncertain. Therefore it is advisable to generalize the results only with caution.

3.) Greenland. The Equip Sermia in Westgreenland in 70° n.lat. shows at the tongue-end a small retreat.

Development in future time.

It is to suggest that in the coming years especially for the next Geophysical Year more glaciers will be surveyed precisely and that measurements will be repeated for glaciological purposes. Therefore some experiences may be *communicated* especially such ones made during the last three cool and wet summers.

In cool and wet summers it is very difficult and time consuming to obtain suitable glacier surveys with an accuracy of 0,50 m in the firn regions. Also the borders between firn and solid terrain cannot be fixed clearly enough. Due to repeated fall of fresh snow the firn surfaces are absolutely white during the whole summer. They are also during temporarily good weather without seizable details. The stereophotogrammetric results are therefore incomplete and insufficient. In order to bridge over these insufficient results it is of use to survey a small «test» glacier with particular accuracy. On small glaciers the surveying distances are also small. It is possible even a short time after fresh snowfalls to perform succesfully stereophotogrammetric surveys as experience has proved. The author has used as test glacier for the northern East alps the « Nördlicher Schneeferner » near the Zugspitze in the Wetterstein Alps. It is only 0,3 qkm in size but it is a glacier still entirely active. One can reach it by a mountain-railway. On the middle of it near the snowline a gauge has been established. Position and height of this gauge is fixed every year. One can obtain by that the whole accumulation of the year as well as the horizontal and vertical movement. A detailed report about the importance of such a test glacier in connection with the other glaciers surveyed will be given later.

There is a further question how often the precise measurement of several glaciers has to be repeated and how many glaciers have to be investigated. It is of importance that the efforts and expenses neither of the fieldwork nor of plotting and working out should be too great and that the results can be shown in a clear form. According to the experience of the author about measurements at glaciers of the eastern alps it should be sufficient for this part of the alps to repeat the measurement at 10 typical glaciers every five years. But it is necessary to survey annually one or two small favourably located test glaciers.

INVESTIGATIONS ON GRINNELL AND SPERRY GLACIERS, GLACIER NATIONAL PARK, MONTANA (*)

ARTHUR JOHNSON
United States Geological Survey Washington, D. C.

Abstract

Information on the extent and characteristics of the Grinnell and Sperry Glaciers was first recorded on the Chief Mountain topographic quadrangle mapped in 1900-1902. Annual determinations of terminal recession were initiated in 1931. Profiles to show changes in surface elevations have been measured annually on Grinnell Glacier starting in 1950. Similar measurements were made on Sperry Glacier in 1949, 1950, 1952, and 1956. Data on rates of movement have been obtained. Precipitation on and runoff from Grinnell Glacier and its enclosing cirque have been measured since 1949. A map of each glacier was compiled in 1952 from aerial photographs taken in 1950. The information now available indicates that the rapid rate of shrinkage that occurred during the 1920's and 1930's has appreciably decreased since the late 1940's. The variations on the Grinnell Glacier since 1950 have been within such narrow limits that the glacier can be considered as having been in a state of balance during this period. The Sperry Glacier has shown a continued terminal recession, but at a decreasing rate.

Introduction

The Grinnell and Sperry Glaciers, at latitude 48''45' and 48''38' respectively, are now presumably the two largest glaciers in Glacier National Park. They are 17 and 25 miles south of the Canadian border, nine miles apart, and located on opposite sides of the Continental Divide. The Grinnell Glacier is in the Hudson Bay drainage and the Sperry Glacier is in the Pacific Ocean drainage.

The Grinnell Glacier was named in honor of George Bird Grinnell who was one of the first to explore the mountains in that area. His first trip to the glacier was in 1887 and several photographs taken at that time are in the files of the Park Naturalist at Glacier National Park. It is believed that the pictures taken that year were not only the first to be taken of the Grinnell Glacier but of any glacier in the area that is now Glacier National Park. A comparison of these with subsequent pictures provides a measure of the changes that have occurred in the intervening periods.

The Sperry Glacier was named for Dr. Lyman B. Sperry, of Oberlin College. He camped in Avalanche Basin in 1894 and deduced from the milkiness of the water flowing into Avalanche Lake that it came from a glacier. In 1895 he succeeded in scaling the cliffs at the head of the basin and reached a point from which the glacier could be seen. Albert L. Sperry (1938, p. 51), nephew of Dr. Sperry, and a member of his party describe the impressions gained from their first view of the glacier as follows:

«While standing upon that peak overlooking the terrain above the rim wall, we got the thrill of thrills, for there lay the glacier, shriveled and shrunken from its former size, almost senile, with its back against the mountain walls to the east of it, putting up its last fight for life. It was still what seemed to be a lusty giant, but

(*) Publication authorized by the Director, U. S. Geological Survey.

it was dying, dying, dying every score of years and as it receded, it was spuing at its mouth the accumulations buried within its bosom for centuries.»

The glacier was first actually reached in 1897 by Dr. Sperry and his party. The route followed that year took them near the present site of the Sperry Chalet and through the saddle between Edwards and Gunsight Mountains, the same genera route of approach now in use. Several pictures taken on this trip were published in a book entitled Avalanche (Sperry, 1938). They are of considerable general interest but are not as useful for comparative studies as the early pictures of the Grinell Glacier. One of these pictures shows clearly a part of the teminal moraine which is now a prominent topographic feature 1500-2000 feet from the present ice front. The ice front was against the moraine but well below the top of it indicating that recession was in progress at that time.

The first factual data available for the glaciers resulted from the mapping of the Chief Mountain quadrangle by the U. S. Geological Survey (scale 1:125,000, contour interval 100 feet) during the years 1900-1902. The exact year in which the area surrounding the glaciers was mapped is not known.

William C. Alden of the U. S. Geological Survey, made studies in the Glacier National Park area during the summers of 1911, 1912, and 1913. His report (1914) describes the glaciers and discusses the glacial phenomena observed during the course of the above studies.

George C. Ruhle, Park Naturalist, Glacier National Park, started terminal recession measurements of the Grinnell and Sperry Glaciers along with others in 1931. These measurements were continued through 1944. Data on terminal recession are among those which have been obtained since 1945 by the Park Service and the U. S. Geological Survey working together.

J. L. Dyson mapped the Grinnell Glacier in 1937 and the Sperry Glacier in 1938 and remapped the Grinnell Glacier and lower portions of the Sperry Glacier in 1946. He has published several reports based on the results of his mapping and related studies which will be referred to later.

Maps of the Grinnell and Sperry Glaciers have been compiled by the U. S. Geological Survey from aerial photographs taken in 1950.

The results of the various observations and studies are described for each glacier.

Grinnell Glacier

Area and Recession

The Grinnell Glacier at the time of the mapping of the Chief Mountain quadrangle had an area of 570 acres. At that time the glacier consisted of two parts, a very steep upper part lying on a bench in the western part of the cirque and a lower larger part — the main mass of the glacier. These two parts were connected by a narrow tongue of ice. This disappeared some time between 1911 and 1929 (Dyson and Gibson 1939, p 683) leaving two independent ice bodies. The lower one is considered herein to be the Grinnell Glacier. The area of the glacier, based on the mapping by Dyson was 328 acres in 1937 and 280 acres in 1946. The area in 1950 as determined from the U. S. Geological Survey map was 257 acres. Dyson (1940, p. 508) estimated that the shrinkage of the glacier during the 40 year period 1897-1937 may have been more than 50 percent of its volume.

Recession measurements of the Grinnell Glacier were started by the Park Naturalist in 1932. These were made from marked control points to several points along the ice front and the average increase in the distance from the control points to the ice front was taken as the terminal recession since the previous measurement.

The recession from 1932 until 1937, the year the glacier was first mapped by Dyson, totaled 165 feet or an average of 33 feet per year (Beatty and Robinson, 1948). A comparison of the 1937 and 1946 maps by Dyson (1948, p 101) shows that there was an average recession of 318 feet along 6540 feet of the glacier front during the intervening period or an average of 35 feet per year.

In 1945 a new method of measuring terminal recession was initiated by the National Park Service and the Geological Survey working cooperatively. Several points were permanently marked on the rock ledges in front of the glacier. These were used as instrument stations and the terminus mapped by plane table methods. A comparison was made for a half-mile section of the ice front, or terminus, extending soootheastward from the lake at the northern border of the glacier. The front in the lake was excluded since changes there due to sections breaking away into the lake would not necessarily be indicative of the true recession. The results for the half-mile section are as follows:

	Period	Recession (feet)	Average annual recession (feet)	Total recession since 1937 (feet)
a/	1937-1945	270	33.8	270
	1945-1947	29	14.5	299
	1947-1950	46	15.3	345
	1950-1952	5	2.5	349

a/ 1937 position of terminus from map by Dyson.

Tje recession was not uniform along the half-mile section. A 500-foot portion of the front was much the same in 1945, 1947, 1950 and 1952. The ice front has not been mapped since 1952 but pronounced changes since then are not apparent. The apparent recession for the 1950-1952 period should not be taken too literally since the irregularities in the topography of the bed rock along the ice front have a considerable influence on its actual position. Even allowing for possible inconsistencies. the foregoing table indicates that there was a pronounced decrease in the rate of recession starting about 1945 and that the recession since 1950 has been negligible.

Changes in Surface Elevation

Since 1950 two profiles, originating from the same point, and making an angle of 52° to each other, have been measured. For purposes of analysis and comparison these profiles have been broken into segments and the mean elevation computed for each year of observation. The results of this analysis are shown in the following tables.

PROFILE 1

Mean Elevation, Feet

Date	Distances from initial point, feet			
	100-500	500-1000	100-1500	1500-2000
Sept. 14, 1950	6463.5	6510.3	—	—
Aug. 22, 1952	6463.6	6510.0	6529.5	—
Sept. 4, 1953	6460.2	6505.5	6523.6	—
Sept. 27, 1954	6461.1	6505.8	—	—
Sept. 8, 1955	6462.0	6505.1	6523.9	—
Aug. 30, 1956	6462.6	6504.4	6562.9	6515.4

PROFILE 2

Mean Elevation, Feet

Date	Distances from initial point, feet				
	100-500	500-1000	1000-1500	1500-2000	2000-2500
Sept. 14, 1950	6460.1	6523.3	6564.8	—	—
Aug. 22, 1952	6460.3	6522.6	6563.8	6604.8	—
Sept. 4, 1953	6458.4	6519.5	—	—	—
Sept. 27, 1954	6459.5	6522.0	6564.4	—	—
Sept. 6, 1955	6460.6	6521.8	6563.9	—	—
Aug. 30, 1956	6461.7	6521.6	6563.8	6604.6	6659.9

The results in the above tables can be summarized as follows: The 1950 and 1952 values for profile 1 were essentially the same. There was an appreciable decrease from 1952 to 1953, varying from three to six feet. Since 1953 the segment from 100 to 500 feet from the initial point has indicated a gradual increase totaling slightly over two feet whereas the segments from 500 to 1000 and 1000 to 1500 feet have indicated a net lowering of about one foot. The lower elevation for the segment from 1500 to 2000 feet as compared with the segment from 1000 to 1500 feet is considered due to the abrasive action of the stream which falls from the cirque wall onto the glacier in this area.

The conditions along profile 2 have remained much the same during the 1950-1956 period as shown by the foregoing table. The 1956 and 1952 profiles were essentially the same throughout with the exception of a slight increase in the first 400 feet from the initial point shown by the 1956 measurements. The tabular data show a lowering from 1952 to 1953 of from two to three feet, a change similar to that noted for profile 1. This lowering tendency has been overcome in subsequent years and the 1956 values are practically the same as the 1952 values.

In brief, the data in the above tables indicate that at profile 1 there has been a lowering of the surface elevation sinced 1952 in amounts up to as much as six feet, whereas for profile 2 the changes in the surface elevations of the glacier have been within such narrow limits that it can be considered as being essentially the same throughout the six-year period 1950-1956.

It is of interest to note that there was no essential difference in profile 2 between 1952 and 1956 whereas profile 1 showed a definite lowering during that period. The fact that the latter is nearest the lake at the northern end of the glacier and, as will be explained later, appears to be the actual terminus, may have some significance. This may also indicate a condition similar to that observed on the Sperry Glacier, viz., that the section adjacent to the terminus is lowering whereas the upper part remains much the same. The differences exhibited by these two profiles should be carefully watched in future years.

Movement

The first measurements of movement were made by Alden (1923, p 268) during the four-day period August 26-30, 1920, a period of chilly, rainy, and snowy weather. Four markers were set in the frontal edge of the glacier. The movement ranged from one inch to 4 3/4 inches during the four-day period. These measurements were characterized by Alden as too crude to form a basis for estimating the average daily or annual rate of advance of the ice. They do indicate, however, that the rate of movement was rather slow.

Dyson and Gibson (1939, p 689) estimated the rate of movement by considering the relation between statification lines, assumed to be the surface expression of boundary planes between successive seasonal accumulations of ice, and distance across the glacier. Careful study of photographs with a hand lens identified 60 bands from the south cirque wall to a point on the east front, a distance of 1800 feet, giving an annual movement of 30 feet per year.

Data on movement have been obtained since 1947 by relocating marked boulders from year to year. Three boulders marked in 1947, two in 1950, and there in 1952 have all shown a movement of from 35 to 40 feet per year in parallel directions. This direction has been generally north or slightly east of north whereas the ice edge which had previously been considered the terminus or front has a northwest-southeast direction. It now appears that this is more a lateral boundary or edge rather than the terminus whereas the actual terminus is at the southern edge of the lake now forming the northern border of the glacier. It is of interest to note that the general direction of movement indicated by the marked boulders is much the same as the direction of the two long, narrow screen areas shown on the 1950 U. S. Geological Survey map. It is also of interest to note the fraily close agreement between the rate of movement since 1947 with that determined by Dyson and Gibson (1939, p 689).

Precipitation and Runoff

In order to attain a better understanding of the relation and develop possible correlations between precipitation and runoff, and glacier variations the investigations at Grinnell Glacier have included precipitation and runoff records. In 1949 the U. S. Weather Bureau, cooperating with the Park Service, installed a storage precipitation gage at the end of the horse trail, within one-half mile of the glacier. During the same year the Helena office of the Geological Survey installed a gaging station at the outlet of Grinnell Lake to measure the runoff from the Grinnell Glacier and its enclosing cirque. This gaging station is nearer to a glacier than any other gaging station in the United States. A continuous temperature record has been obtained at this

gaging station since 1951 and summer precipitation measurements were started in 1956. A second storage precipitation gage was installed jointly by the Weather Bureau and the Park Service in 1955 about 3,000 feet south-southeast of the 1949 installation. These gages are now designated a Grinnell No. 1 and Grinnell No. 2 in Weather Bureau records.

Dightman (1956, p 329-332) has analyzed the temperature and precipitation records for three stations, Kalispell, West Glacier, and Fortine, which are within 60 miles of the Grinnell Glacier. The longest record is for Kalispell which extends back to 1897. The ten year moving average for temperature at this station shows an increase from about 1905 to 1915, very little change from 1915 to the mid 1920's, a definite increase from then until the early 1940's, and a general decrease thereafter. The ten year moving average for precipitation shows a pronounced decrease from 1915 to 1925, a slight increase from 1925 to the early 1940's and a marked increase through 1951. Records for West Glacier and Fortine, starting in the 1930's show variation similar to those recorded at Kalispell.

The pronounced shrinkage and recession of the Grinnell Glacier which occurred during the 1920's and 1930's coincides with the higher than average temperatures and decreased precipitation occurring during that period. The approximately stable condition of the glacier observed since about the mid 1940's coincides with the lower average temperatures and greater precipitation that has been evident since the early 1940's. If the general trend of temperature and precipitation that has existed since the early 1940's continues for several more years it seems logical to assume that the glacier may begin to increase in size and volume.

Sperry Glacier

Area and Recession

The Sperry Glacier, now considered to be the largest single ice-mass in Glacier National Park, had an area of 840 acres at the time the Chief Mountain quadrangle was mapped. The area in 1938 was shwon by Dyson's map to be 390 acres, and in 1950 it was 298 acres according to the U. S. Geological Survey map.

Dyson (1948, p 97) from a study of 1913 photographs estimated that the average recession of the terminus from that year until 1938 was 1533 feet measured along 5700 feet of the front, or an annual average of 61 feet.

Markers for recession measurement were set by the Park Naturalist in 1931. These were not found in 1932 and were presumed to have been covered with snow from the previous winter. No data were obtained for the years 1933 and 1934. New markers were set in 1935 and apparently the 1931 markers were not found as no data are given for the 1931-1935 period. The recession for the 10-year period 1935-1945 was 641 feet or an average of 64 feet per year. This agrees closely with the annual value for the 1913-1938 period mentioned above. The mapping in 1938 did not tie in the markers used for recession measurements so a correlation between the recession measurements and the mapping is not possible.

The method of determining terminal recession by plane table mapping as described for the Grinnell Glacier was also initiated for the Sperry Glacier in 1945. The terminus was remapped in 1947, 1948, 1949, 1950, 1952, and 1956. The total recession and annual rate for a half-mile section near the center of the front for selected periods is shown below.

Recession of terminus, in feet, of central half-mile portion

	Period	Recession for period	Annual rate
a/	1938-1945	351	50
	1945-1950	177	35
	1950-1956	85	14

a/ Position of 1938 terminus from map by J. L. Dyson.

The recession along the half-mile section considered was not uniform. During the 1938-1945 period the change varied from a maximum of 600 feet to a minimum of less than 100 feet. Similarly, during the 1950-1956 period the change varied from more than 500 feet to less than 100 feet. The terminus at the time of mapping in 1956 was still covered in some places with snow from the previous winter and the exact ice front could not be clearly defined. Consequently, the front as mapped may, in places, show it to be farther down valley than is actually the case. The total and annual recession for the 1950-1956 period may therefore be somewhat greater than shown but the amount is not believed to be appreciable. Even allowing for some error in the above table for the 1950-1956 period the data indicate that a pronounced decrease in the rate of recession has occurred during the period of record.

The maximum recession has occurred at about the midpoint of the front. In 1938 the front was farthest forward at this point. The front of the ice farthest up-valley in 1956 was opposite the most forward point in 1938. It appears that the terminus is gradually changing from a convex front as shown on the 1938 map to a concave front as indicated by the 1956 mapping. Such a change is to be expected in a receding glacier.

Changes in Surface Elevation

Based on studies of the Nisqually Glacier on Mt. Rainier, Washington (Johnson, 1953) terminal recession does not necessarily indicate a shrinkage or lowering of the surface of the entire glacier. In order to obtain quantitative data on surface changes of the Sperry Glacier a profile was measured across the glacier approximately midway between the terminus and upper boundary in 1949 and remeasured in 1950, 1952, and 1956. A longitudinal profile at about right angles to this cross-profile and about two-thirds of the distance across the glacier from the left or west edge was also established in 1949 and remeasured in 1950, 1952, and 1956.

From a comparison of the profiles for the years of record the following general statements can be made.

The results of the measurement along the cross-profile in 1950 were practically the same as those observed in 1949. The 1952 profile was above that for 1950 across the entire glacier, in places as much as ten feet. These maximum differences were in the side portions with little change evidenced in the central section. Similarly, a comparison of the 1956 and 1950 profiles shows that the 1956 surface was well above that for 1950 from the left or west edge for a distance of 1600 feet with maximum observed differences up to 15 feet; across the next 500 feet there was very little difference between the 1956, 1952, and 1950 surfaces; and from there on to the east edge, a distance of about 1000 feet, the 1956 surface was above that of 1950 by as much as

15 feet. The lowest part of the profile is about 1700 feet from the west edge of the glacier and it is of interest to note that for several hundred feet in this area there was no appreciable difference between the profiles in 1950, 1952 and 1956.

In comparing the 1952 and 1956 results with 1950 it must be kept in mind that the observations in 1950 were made about a month later than those in 1952 and 1956. Even allowing for the ablation during that month interval it would appear that the 1956 profile would be considerably above the 1950 profile with the exception of the central portion already mentioned. Whether or not this increase in surface elevation represents the beginning of a build-up in this area of the glacier or merely a variation in the residual snowfalls from year to year is a point that can only be determined by measurements in future years.

The changes which have been disclosed by measurements of the longitudinal profile are particularly noteworthy. The lower portion of the glacier has continued to decrease in surface elevation whereas the upper portion shows a slight increase or build-up. The point of transition is approximately 2,000 feet along the line of the profile from the 1956 terminus of the glacier and near the point of intersection of the longitudinal profile and the cross profile. The 1956 surface near the terminus was as much as 35 feet below the 1952 surface and as much as 40 feet below the 1950 surface. At a point about 750 feet back from the 1956 terminus the 1950, 1952, and 1956 profiles intersect. The section immediately upglacier from this point of intersection exhibits an interesting phenomena. The ridge paralleling the front, which has been a prominent feature of this glacier for many years, has been increasing in elevation and moving upglacier. The crest of this ridge was 16 feet higher in 1956 than in 1950 and about 60 feet farther upglacier. A change amounting to about an eight-foot increase in elevation and a 30-40 foot movement upglacier was noted to have occurred between 1952 and 1956. Similarly, the trough or depression on the upper side of this ridge had likewise been moving upglacier. The low part of this trough was about 100 feet farther upglacier in 1956 than it was in 1950 and three to four feet higher in elevation. The change in the position of the low point from 1952 to 1956 was about 75 feet and the low part of the depression in 1956 was two to three feet lower than in 1952. Continuing up the glacier the 1956 surface was definitely below that of 1952 and 1950 for the next six or seven hundred feet. From that point on the 1956 surface was above that of 1952 and 1950 by varying amounts. The maximum difference between the 1956 and 1952 surfaces was about five feet whereas the difference between the 1956 and 1950 surfaces was from 12 to 15 feet. As previously stated any comparison between 1956, 1952, and 1950 must take into account the difference in time of the year at which the measurements were made. Allowing for this time difference in observations, the data indicate that the upper part of the glacier, if not actually increasing in surface elevation, is at least holding its own whereas the lower portion is definitely decreasing in elevation.

As a further means of comparison the mean elevation of segments of the profiles were determined for the years of record. These results are shown in the following tables and are self-explanatory.

CROSS PROFILE

Mean elevation, feet

Date	Distances from initial point, feet *a/*			
	100-1100	1100-2100	2100-3100	100-3100
Aug. 30, 1949	7598.9	7535.1	7593.6	7575.9
Sept. 19, 1950	7597.9	7534.5	7594.3	7575.5
Aug. 19, 1952	7603.5	7539.0	7600.4	7581.0
Aug. 23, 1956	7607.0	7539.9	7603.5	7583.4

a/ Initial point is marker on ledge rock within 100 feet of west edge of glacier.

LONGITUDINAL PROFILE

Mean elevation, feet

Date	Distances from initial point, feet *a/*						
	Downstream from initial point feet				Upstream from initial point feet		
	1500-1700	1000-1500	500-1000	0-500	0-500	500-800	500-1000
Aug. 30, 1949	—	—	7449.2	7492.4	7569.9	—	—
Sept. 19, 1950	7331.6	7398.2	7447.6	7491.2	7570.7	7663.9	—
Aug. 19, 1952	7324.3	7397.0	7452.6	7493.5	7576.0	7671.8	—
Aug. 23, 24, 1956	7290.2	7377.6	7453.8	7888.2	7577.0	7676.7	7716.1

a/ Initial point for the longitudinal profile is on the cross profile and 1967 feet from reference point therefor.

Movement

Alden (1914, p. 15) made several observations on rate of movement in a cave at the front of the glacier for a 24-hour period, August 16.17, 1933. The maximum rate observed was 1 inch in 24 hours.

In 1949, at the time the profile measurements were initiated, four rocks were marked and located. Two of these were found and relocated in 1956. Each showed a movement of about 80 feet for the seven-year period and a lowering in elevation of between seven and eight feet. These rocks are located about 1800 feet from the 1956 terminus and about 1800 feet from the left edge. The foregoing observations indicate a rate of movement of approximately 10 feet per year in this area of the glacier. Considering the shape of the glacier it seems possible that a slightly greater rate of movement might occur about midway between the rocks observed and the west edge. It would be advisable to obtain data on movement in this area as a part

of future investigations. The data now available indicate that the Sperry Glacier can well be considered as having a slow rate of movement.

Summary

From the data now available the following generalizations can be made. The very rapid rate of recession and shrinkage of the Grinnell and Sperry Glaciers that occurred during the early 1900's and particularly during the 1920's and 1930's has decreased appreciably. This change has been quite evident since the early or mid-1940's. The observations on the Grinnell Glacier indicate that it can be considered as having been essentially in a state of balance for the past decade. The Sperry Glacier has shown a continual recession of the terminus but at a much lesser rate since about 1945. Based on observations since 1949 the lower portion of the glacier has continued to decrease in surface elevation along with the terminal recession whereas the surface elevation of the upper portion has remained nearly the same with some indication of a slight build-up or increase.

The change in trend occurring in the 1940's agrees with the change in trend of increased precipitation and decreased temperatures indicated by weather records.

Glaciers have been described as «historians of climate» as well as «integrators of climate.» The relationship between glacier variations and climatic changes must be viewed over a long period rather than on a year-to-year basis. The data that have now been accumulated for the Grinnell and Sperry Glaciers, along with studies planned for the future, will contribute to a better understanding of the relationship between glacier variations and climatic changes.

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LES INVESTIGATIONS GLACIOLOGIQUES EN L'URSS

Prof. G. A. AVSIUK,

Summary

I. Introduction

1. Glaciological investigations during the International Geophysical Year (IGY), 1957-1958, will yield numerous fresh data which will provide for a further progress of glaciology.

2. The information on the state of glaciological investigations before the IGY which is of a definite scientific interest and is directly connected with the IGY operations.

3. The report includes: the information on the glaciological investigations in the USSR before the IGY and during the IGY, as well as some data on peculiarities of glacial morphology for various ice regions of our country.

4. In the recent time many scientists regard glaciology as a science about all ice formations on the Earth. However, this report is confined only to the investigation of glaciers proper.

II. Progress in glaciology in the USSR before the IGY

1. Glaciological investigations in the USSR before the Second Polar Year (IPY), 1932—1933. A rather late beginning of the Russian glaciological investigations on glaciers. A subordinate role of glaciological operations at the first stage of their development. The intensification of glaciological investigations before the Second IPY and their general characteristics.

2. Glaciological investigations during the Second IPY (1932—1933) and the so-called «glacier expeditions» within that period. Expeditions to the Arctic, Near-Polar Urals, Caucasus, Altai, Tien-Shan, Pamirs. The glacial operations in the Second IPY as a big step in the advance of glaciology of the Soviet Union.

3. Glaciological investigations in the interim period between the Second IPY (1932—1933) and the IGY (1957—1958). Glaciological work in the Tien-Shan, Pamirs, Altai, Caucasus and the Arctic. The discovery of new regions of modern ice formation in the Upper Indigirka and Polar Urals. The organization of stationary glaciological investigations in the Antarctic. The beginning of glaciological operations in the Antarctic and general characteristics of this period.

III. Chief results of glaciological investigations in the USSR by the beginning of the IGY

1. The characteristics of the modern progress of ice formation on the USSR territory as a whole and on separate glacial regions: Franz Josef Land, Novaya Zemlya, Northern Land, the Island of Eastern Arctic Seas, Polar and Near-Polar Urals, Taimyr peninsula, the Indigirka Reservoir, Kamchatka, the Caucasus, Altai, Tien-Shan, the Pamirs.

2. Main peculiarities of the modern ice formation in various glacial regions, mostly, as evidence of the influence of continental climatic trends.

3. The modern ice formation, its probable stability, trends of the ice formation. evolution. Probable causes of the modern trends in ice formation.

4. Notions on hionosphere, energy and intensity of ice formation.

5. The study of the problems on: nourishment, formation and structure of ice; temperature and different processes of glaciers under physical and geographical conditions.

6. Basic glacier types according to the ice formation and temperature regimen.

7. Zones of glaciers and the dependence of their activity on physical and geographical conditions.

8. Basic laws of the modern distribution of glaciers with different types of ice formations.

9. The treatment of problems on structural glaciology, i. e. a science on ice rocks and their formation processes.

10. The treatment of problems on the interaction of climate and ice formation.
11. The treatment of problems on the influence of climate and relief upon the progress of ice formation.
12. The treatment of problems on the use of glaciers as water sources for irrigation and power stations.
13. The directions of further glaciological investigations.
14. The list of the main works on glaciology, published in the USSR for the last five years, with their short characteristics.

IV. The basic features of glacial morphology and peculiarities of glacier processes in various regions of modern ice formation in the USSR

1. General conceptions as a basis for further study of the morphology and processes of glaciers.
2. The notion «glacier type» used in the characteristics of glaciers' morphology.
3. Rejection of the classical «alpine» formula (the ratio between the square of the nevé slope and that of the ice-tongue as an index of «normality» of this or that glacier) considered to be universally just.
4. The conditionality of the notion «nevé line».
5. The main features of glacial morphology and processes; data on the areas of modern ice formation from evidence of separate regions of the USSR.
6. Data on the total area of modern ice formation in the USSR.

V. Glaciological operations during the IGY

1. The main objectives of glaciological investigations in the IGY and a summary of their problems' in the USSR.
2. The IGY investigations as a natural continuation of the earlier glaciological operations in the USSR.
3. The IGY glaciological stations in the USSR. The principle according to which they were selected.
4. Special significance of glaciological investigations in the Antarctic.
5. Some information on the progress of glaciological investigations under the IGY program.

I. Introduction

Les investigations étendues, entreprises actuellement selon le programme de l'Année Géophysique Internationale de 1957-1958, donneront un nouveau matériel immense qui embrasse les phénomènes glaciologiques sous tous leurs aspects. Ceci crée des conditions favorables pour une importante évolution ultérieure de la glaciologie.

C'est pourquoi une information qui éclaire l'état des investigations glaciologiques avant le commencement de l'Année Géophysique Internationale (l'A.G.I.) porte un grand intérêt scientifique. Une courte information de ce genre caractérisant en traits généraux l'état de ces études dans l'Union Soviétique avant le commencement de l'A.G.I. fait en somme le contenu de ma communication. Naturellement, on y trouver aussi certains renseignements relatifs aux particularités essentielles des diverses régions de congélation contemporaine dans notre pays.

Tout dernièrement, beaucoup de savants estimèrent la glaciologie comme une science qui étudie toutes les formes de glace sur la Terre. En ne discutant toutefois point cette compréhension étendue de la glaciologie, nous nous limitons néanmoins dans notre exposé à l'examen d'une série de questions qui touchent rien que les glaciers contemporains comme tels.

II. Développement des travaux glaciologiques en URSS avant l'A.G.I.

La glaciologie est une science relativement jeune; A peine deux siècles sont écoulés depuis que l'on a effectué les premières observations scientifiques sur les glaciers alpins.

En Russie l'étude immédiate des glaciers commença beaucoup plus tard, — seulement durant les années soixante du siècle précédent. De cette manière les investigations en glaciologie de notre pays n'atteignent pas encore une centaine d'années.

Le développement des investigations glaciologiques dans les régions glaciaires de notre pays peut être divisé en trois périodes principales : la première d'entre elles embrasse le temps entre les ans soixante du siècle précédant et le commencement de la 2-ème Année Polaire Internationale (A.P.I.); la deuxième comprend la Seconde Année Polaire Internationale, c'est-à-dire les années 1932 et 1933; et la troisième période embrasse le laps de temps après la seconde Année Polaire Internationale jusqu'au commencement de l'Année Géophysique Internationale (1957-1958).

La première période se caractérise en entier par l'exécution de travaux glaciologiques descriptifs à base desquels l'on pouvait se faire une première impression de la congélation contemporaine dans notre pays. Ce furent presque exclusivement les « langues » des glaciers que l'on étudia. On ne visita point de règle les parties supérieures des glaciers et leurs descriptions furent approximatives. On n'exécuta pas non plus des levés précis pour les glaciers; c'est ainsi que les dimensions et les surfaces de ces derniers furent estimées d'une manière approximative. L'on ne fit point non plus d'observations systématiques sur les processus naturels propres aux gletschers. Ordinairement l'on accordait à la caractéristique des types de glaciers et aux processus qui ont lieu sur ces derniers relativement peu d'attention; cela se faisait principalement par voie de conclusions spéculatives, fondées souvent sur des conceptions empruntées mécaniquement à la glaciologie alpine.

L'étude des glaciers portait durant cette période un caractère en général accidentel. Les observations glaciologiques se faisaient accessoirement avec les investigations géologiques, parfois botaniques ou zoogéographiques et elles jouaient un rôle secondaire. Les observations glaciologiques etaient généralement exécutées non pas par des glaciologues, mais par des personnes ayant des spécialités les plus diverses.

Ces investigations étaient d'un ordre géologique ou géomorphologique fortement exprimé et se rapportaient principalement à l'étude des activités d'accumulation et d'érosion des congélations actuelles et anciennes. On peut expliquer particulièrement de par cela la circonstance suivante — l'attention principale était dirigée vers l'étude des langues glaciairer. Il est vrai que la glaciologie alpine se développait aussi généralement dans la même direction géologo-géomorphologique.

Il est nécessaire de noter, que nonobstant les conditions peu favorables indiquées du développement de la glaciologie russe durant la première période, l'accumulation des données sur les glaciers de notre pays se faisait néanmoins assez rapidement.

Les éminents savants russes de cette époque, comme par exemple I. V. Mouschketov, A. I. Voieykov, P. A. Krapotkin, B. A. Fedchenko et autres se rendaient bien compte de l'imperfection de tels travaux glaciologiques. Ils indiquèrent à plusieurs reprises dans leurs œuvres la nécessité d'une approche plus approfondie et plus étendue pour l'étude des glaciers. Ils indiquèrent aussi l'inadmissibilité de reporter mécaniquement les lois gouvernant les glaciers alpins sur n'importe quels glaciers dans d'autres régions. Grâce à une compréhension juste et étendue des phénomènes glaciaires et malgré l'imperfection et le peu de matériel fondé sur les faits étant à leur disposition, ces savants créèrent des grandes œuvres classiques qui eurent une influence prononcée sur le développement de la glaciologie et qui n'ont point perdu leur importance à notre époque. Il est suffisant, par exemple, de nommer l'œuvre de P. A. Kropotkine, intitulée « Investigations sur la période Glaciaire », ou bien

celle de A. V. Voeyikov « Conditions climatiques des phénomènes glaciaires, présents et passés. »

Après la Grande Révolution Socialiste d'Octobre, le rythme des investigations glaciologiques s'est vivement accéléré dans l'Union Soviétique. Une évolution particulière de ces investigations se fit sentir en premier lieu dans les régions montagneuse de l'Asie Centrale. Ces investigations commencèrent à gagner une importance indépendante et à embrasser beaucoup de questions d'une grande étendue, nonobstant que ces dernières conservaient encore une forte direction géologique et géomorphologique.

Vers les années trente de notre siècle les investigations glaciologiques s'intensifièrent particulièrement et en plus de l'ancienne direction géologique et géomorphologique il se développa une nouvelle approche pour l'étude des glaciers comme phénmènes hydro-météorologiques. Il se traça de même un écart décisif des canons de la glaciologie alpine; cet écart s'éprima en premier lieu par ce que l'on commença à tenir compte en étudiant les phénomènes glaciaires des différentes particularités dans les conditions physico-géographiques des régions glaciaires variées.

Cette période se distingue par une plus grande manifestation de l'approche géographique complexe pour l'étude des glaciers. Avec cela, la fin des années vingt est caractérisée par la découverte d'une grande quantité de nouveaux glaciers dans l'Altaï, le Thian-Chan, le Pamir et d'une nouvelle région glaciaire dans l'Oural Prépolaire. C'est à cette période en particulier que l'on a étudié d'une manière complète un des plus grands glaciers montagneux du monde, celui qui porte le nom de Fedchenko.

Une place à part est occupée dans l'évolution de la glaciologie soviétique par la Deuxième Année Polaire Internationale de 1932-1933, qui fait en même temps la seconde période du développement de la glaciologie dans notre pays.

Notons que la proposition d'élargir le programme des investigations glaciologiques durant la 2-ème Année Polaire Internationale (2 AGI), c'est-à-dire d'y inclure non seulement les glaciers des pays polaires, mais encore ceux des régions montagneuses de congélation contemporaine dans les latitudes modérées fut faite par le Comité Soviétique pour l'exécution de la 2 API. Ce programme fut exécuté ultérieurement seulement dans l'Union Soviétique.

Vers le commencement des travaux de la 2 API il s'affermit dans l'Union Soviétique parmi la plupart des glaciologues l'idée de la nécessité d'une approche géographique complexe pour l'étude des glaciers. Une attention particulière fut apportée à l'étude des glaciers comme phénomènes hydrométéorologiques ; ceci reflète la direction pratique des investigations glaciologiques de ces temps-la. Ce dernier aspect des investigations glaciologiques est particulièrement caractéristique pour les travaux de la 2-ème API.

On crée pour l'exécution des investigations glaciologiques de la 2 API des expéditions glaciaires spéciales. Les travaux de ces expéditions ont embrassé simultanément presque toutes les régions de congélation connues alors sur le territoire de l'URSS : le Caucase, l'Altaï, le Pamir, le Thian-Chan, l'Oural Prépolaire et les îles Arctiques.

Durant les travaux des expéditions, on a organisé sur les glaciers des stations météorologiques temporaires; on a pour la première fois fait des mesurements multiples de l'ablation, des vitesses du mouvement de la glace; on a accordé beaucoup d'attention à l'étude de la morphologie des gletchers, ainsi qu'aux discriptions de la structure des glaces des différentes formations glaciaires sur la surface des glaciers, etc. Sur les courants d'eau qui prennent naissance dans les glaciers, on a exécuté des travaux hydrométriques. Maints glaciers, particulièrement leurs parties terminales, ont été levées à l'aide d'instruments. Le volume du matériel obtenu par les expéditions glaciaires de la 2-ème API se trouva à vrai dire immense et suffisament divers. Il faut mentionner, il est vrai, qui l'on n'atteignit encore pas dans ces travaux de la

2-ème API l'uniformité voulue et que les procédés mêmes des observations ne touchaient non plus à une grande perfection par suite de quoi les données obtenues avaient une qualité variées.

Les résultats des investigations glaciologiques de la 2-ème API furent reflétés dans six volumes des «Travaux des expéditions glaciaires de la 2-ème API» qui furent publiés en 1935-1936. Le contenu de ces œuvres se distingue par une grande richesse de matériel fourni par les faits et de déductions théoriques qui portent pour la plupart des cas un caractère original. La direction générale des travaux, glaciologiques de la 2-ème API s'y est trouvé suffisamment reflétée : l'approche géographique complexe pour l'étude des phénomènes glaciaires, c'est-à-dire tout le nouveau qui caractérise le développement de la glaciologie soviétique à cette période. Au point de vue présent le défaut principal des travaux de 1932-1933 consiste en l'attention insuffisante à l'étude quantitative des processus glaciaires divers, en la faible application de méthodes géophysiques aux observations.

On peut citer comme œuvre qui généralise toutes les investigations de la 2-ème API le livre de S. V. Kalesnik, intitulé « Régions glaciaires montagneuses de l'URSS », qui fut publié en 1937. L'auteur donne dans cet œuvre une généralisation de toutes les données obtenues pour les glaciers de l'URSS, et un exposé des connaissances contemporaines sur les régions glaciaires de notre pays. Il y reflète avec beaucoup de relief le progrès de la glaciologie soviétique qui fut acquis grâce aux travaux de la 2-ème API et y formule les problèmes des futures investigations glaciologiques. L'importance des travaux de la 2-ème API pour notre glaciologie ne peut être surestimée — ce fut un très grand pas le long de cette voie de progrès.

En résultat des investigations de la 2-ème API la plupart des glaciologues soviétiques s'est affermie dans la conviction qu'un glacier représente un événement physico-géographique original, dont la féconde étude scientifique est possible rien qu'à base d'une approche géographique complexe, accompagnée d'une application obligatoire de méthodes systématiques et précises d'observation.

La troisième étape d'évolution de la glaciologie soviétique comprend la période entre la 2-ème API et le commencement de l'Année Géophysique Internationale, c'est-à-dire le laps de temps entre 1933 et 1957. Cette période est caractérisée par un approfondissement et un élargissement ultérieur des investigations glaciologiques dans toutes les régions glaciaires de notre pays. On inculque de plus en plus dans la pratique des investigations les méthodes systématiques de travail, les procédés quantitatifs précis pour les observations. Les investigations glaciaires se font de règle, avec application de procédés photogrammétriques pour les mesurements, de divers procédés géophysiques pour les observations. Elles se fondent sur les données des analyses cristallographiques, etc. La méthode de balance pour l'étude de la substance et de l'énergie des glaciers est de plus en plus reconnue et est inculquée dans les investigations glaciologiques. Durant les investigations sur les glaciers on y fait des levés précis, on étudie la distribution de la température dans la masse de glace, les processus de formation de la glace, de l'alimentation, de la fonte, de l'écoulement. On accorde une grande attention à l'étude de la structure de la glace, du névé et de la neige; à l'étude du mouvement de la glace dans les glaciers, des conditions climatiques et radiotionnées. En un mot, tout le complexe des processus naturels propres aux glaciers se trouve dans la sphère qui intéresse les glaciologues soviétiques. On exécute des profondes investigations de type stationnaire pour les glaciers arctiques sur l'archipel François-Joseph. Dès 1947, on organisa des observations glaciologiques continues sur le Thian-Chad.

Simultanément l'on apporte beaucoup de précision dans les surfaces de congélation des diverses régions, dans le nombre des glaciers; l'on établit la diversité des types de glaciers dans les différentes régions, l'on dresse des catalogues de glaciers pour une série de régions de congélation. Durant la même période l'on découvre

des nouvelles régions de congélation contemporaine inconnues auparavant dans la partie Nord-Est du pays (sur la haute Indiguirka) et dans l'Oural Polaire.

On étudie les questions portant sur les facteurs de congélation; sur les relations entre la congélation et les conditions orographiques et climatiques; sur les problèmes de l'évolution de la congélation, etc. Une attention particulière est accordée aux particularités zonales du développement des congélations et des processus glaciaires. Enfin l'on exécute des études aux laboratoires pour définir les propriétés de la glace ainsi que l'on fait des études expérimentales en conditions naturelles sur les glaciers mêmes pour accélérer artificiellement leur fonte.

A base de traitement et de généralisation de ces nouvelles données multiples et variées, les glaciologues soviétiques publient beaucoup d'œuvres originales portant sur maints problèmes divers de la glaciologie. Ces travaux ont un caractère non seulement régional qui élucide les lois gouvernant différents phénomènes glaciaires propres à divers glaciers et à diverses régions glaciaires, mais parmi ces œuvres se trouvent de telles qui portent sur le caractère et l'état des glaciers de différentes régions, ainsi que des travaux de théorie générale. Parmi le premier groupe suffisamment multiple de ces travaux on peut citer, par exemple, les suivants : M. V. Tronoff, « Congélation de l'Altaf » (1949); R. D. Zabirov, « Congélation du Pamir » (1955); P. A. Schoumsky, « Congélation contemporaine des Arctiques » (1947); G. A. Avsiuk, « Les glaciers des sommets planes » (1950); « Les glaciers du noeud de montagnes Khan-Tengry », « Certaines données sur la congélation et l'orographie du massif montagneux Ak-Shigriak » (1952). Le second groupe des travaux comprend les œuvres de théorie générale et comporte en premier lieu les livres suivants : « Glaciologie générale » par S. V. Kalesnik (1939); « Problèmes de la glaciologie des montagnes » (1954); «Questions liées à la relation entre le climat et la congélation » (1956) par A. V. Tronoff, « La température de la glace dans les glaciers » par G. A. Avsiuk (1956); « L'énergie de la congélation et la vie des glaciers » (1947); « Principes de l'histoire structurelle des glaces » (1955) par P. A. Schoumsky.

De cette manière la glaciologie soviétique a progressé fortement durant la troisième période de son développement en théorie et elle a accumulé beaucoup de nouveaux faits dans les régions de congélation de l'Union Soviétique. Un des traits les plus caractéristiques de cette période est le perfectionnement prononcé des méthodes et des procédés pour les investigations glaciologiques.

Les glaciologues soviétiques ont apporté à la glaciologie une compréhension plus étendue. Actuellement la plupart des glaciologues soviétiques voient en elle une science qui s'occupe de l'étude de toutes les formes de glace sur la Terre, ainsi que de l'étude des glaces dans la circulation de l'eau sur notre planète, c'est-à-dire de l'étude de la criosphère terrestre. La glaciologie étudie les processus d'accumulation, de transformation et du débit de la glace sur la Terre selon le bilan thermique de cette dernière. Elle fait des investigations sur les lois générales propres aux glaces naturelles, ainsi que sur les lois zonales, les particularités géographiques de la manifestation de ces lois.

Vers le commencement de l'Année Géophysique Internationale la glaciologie soviétique s'est enrichie d'une grande quantité de nouveaux faits; elle s'est fortement avancée en théorie.

Le développement de la partie théorique de glaciologie a mis au premier rang toute une série de grands problèmes, la solution desquels est possible rien qu'à base d'investigations approfondies et étendues. L'application effective des données de la science glaciologique aux besonins pratiques est possible seulement si l'on pose ces investigations avec suffisament d'étendue et si l'on obtient des valeurs quantitatives qui caractérisent les phénomènes glaciaires.

La quatrième étape du développement de la glaciologie dans notre pays sera l'Année Géophysique Internationale. Durant cette période l'on va obtenir des nouvelles

données par les observations systématiques pour presque toutes les régions glaciaires de l'URSS.

Outre cela, notre pays participe durant la période de AGI dans les investigations de l'Antarctide.

III. Résultats essentiels des traveaux glaciologiques dans l'URSS et les directions principales de leur développement ultérieur

Nous avons caractérisé en bref l'évolution de la glaciologie dans l'Union Soviétique vers le début de l'AGI. Dans la partie présente nous citerons certaines données sur les connaissances de la congélation contemporaine du territoire de l'URSS et nous nous arrêterons brièvement sur quelques questions qui méritent une attention plus prononcée.

Actuellement, on connaît toutes les régions de congélation contemporaine sur le territoire de l'URSS et l'on dispose pour toutes ces régions de données qui permettent une description générale des glaciers et une élaboration de leurs catalogues. Néanmoins, le matériel obtenu n'est pas encore mis en ordre pour toutes ces régions, c'est pourquoi nous ne pouvons encore citer les donneés contemporaines pour tout le terrain de congélation et les données portant sur le nombre de glaciers pour telles régions comme le sont, par exemple, le Thian-Chan ou l'Oural Polaire. De cette manière, les connaissances des régions de congélation contemporaine sur le territoire de l'URSS sont actuellement suffisantes sous le point de vue de géographie générale. De beaucoup plus faibles sont les connaissances portant sur les particularités des conditions naturelles pour les régions de congélation et, particulièrement sur la manifestation des processus glaciaires. On a fait des observations systématiques sur les processus glaciaires dans les glaciers isolés rien que pour l'Archipel de François-Joseph et pour le Thian-Chad. En ce que concerne les glaciers des autres régions de congélation, des observations des processus glaciaires se ramènent à des séries peu nombreuses et isolées de telles investigations. Actuellement l'appréciation des formes propres et des particularités de processus glaciaires pour toute la masse des glaciers est fondé principalement sur les résultats de l'étude générale de ces glaciers et ne peut être acceptée comme sûre. D'ailleurs, un tel état est caractéristique non seulement pour l'URSS, mais encore pour la plupart des régions de congélation de toute notre planète. De cette manière, le problème principal des investigations glaciologiques contemporaines, tout comme la précision des données de caractère glaciologique général pour des régions isolées de congélation consiste en l'exécution d'observationsh systématiques stationnaires à l'aide de méthodes précises pour tout le complexe des processus glaciaires. De telles observations vont permettre d'établir par argumentation les lois générales, propres aux phénomènes glaciaires et leurs manifestations zonales dans diverses régions de congélation. Même à l'époque actuelle le nombre relativement minime des observations sur les processus glaciaires, étant à notre disposition, a permis de caractériser non seulement les lois générales propres aux phénomènes glaciaires, mais aussi d'établir la zonalité et la stipulation physicogéographique de l'activité vitale des glaciers, et de donner une image approximative de la distribution spatiale des glaciers de types divers.

Le matériel obtenu a permis, par exemple, à P. A. Schoumsky d'élaborer avec suffisamment de détails une nouvelle branche de la glaciologie : la science structurelle des glaces; une branche qui possède des grandes perspectives pour l'évolution ultérieure de la glaciologie. Ce matériel a permi de même de fonder théoriquement les principales conceptions glaciologiques, telles que l'énergie et l'intensité de la congélation, etc. Ces mêmes observations relativement peu nombreuses sur les processus glaciaires donnent le droit de soutenir que les nombreuses particularités de la congé-

lation contemporaine des diverses régions de notre pays s'expliquent principalement par l'influence des traits continentaux du climat.

Les travaux glaciologiques ultérieurs doivent permettre d'obtenir du matériel pour préciser et pour développer les theses que l'on peut exprimer actuellement rien qu'en général et évidemment ils permettront d'avancer une série de nouvelles lois glaciologiques, qui ne sont pas encore remarquées.

Ultérieurement il faudra accorder une attention spéciale en premier lieu à l'élaboration des problèmes suivants :

1. Est-il possible de donner une définition plus précise pour l'état de congélation, le degré de stabilité de cet état, de la tendance ultérieure à son évolution.
2. Préciser les notions de criosphère, de khionosphère, d'énergie et d'intensité de la congélation.
3. Elaborer les questions portant sur les types divers de formation de la glace, sur l'alimentation, le débit de la substance et sur le régime thermal des masses de glace.
4. Préciser les questions portant sur la stipulation de l'activité vitale des glaciers et des lois de leur disposition spatiale.
5. Elaborer les questions sur la glaciologie structurelle, c'est-à-dire sur la théorie des roches en glace et des lois de leur formation.
6. Préciser les questions portant sur la liaison entre le climat et la congélation.
7. Elaborer les questions sur la liaison entre le climat et le relief et de leur influence sur le développement de la congélation.
8. Etablir les thérories du mouvement de la glace dans les glaciers.
9. Elaborer une classification génétique des glaciers.

Nous supposons que le total des observations obtenues en résultat des investigations exécutées durant l'AGI permettra d'avancer grandement l'élaboration des problèmes glaciologiques qui sont cités plus haut et que cela rehaussera d'une manière importante le niveau de cette science.

IV. Traits principaux de la congélation contemporaine dans les différentes régions du territoire de l'URSS

On a établi actuellement que dans toutes les diverses régions de congélation cette dernière se trouve dans une phase de retraite. Ceci est vrai nonobstant qu'il y a dans l'Union Soviétique des régions de congélation contemporaine qui se trouvent dans des conditons physicogéographiques des plus différentes : à commencer par les Arctiques extrêmes jusqu'à l'Asie Centrale torride; malgré que la congélation dans ces régions diverses occupe des terrains variés, et qu'elles se distinguent par les formes des glaciers et par le caractère des processus glaciaires.

En général, l'état de régression de la congélation contemporaine est caractéristique actuellement pour toute notre planète et à ce qu'il paraît est provoqué par une même raison. Nous supposons que la cause principale de cette regression se trouve dans la hausse des températures.

La congélation contemporaine sur le territoire de l'Union Soviétique se distribue dans les régions suivantes.

A. Sur les îles des Arctiques Soviétiques :

(a) sur l'archipel de François-Joseph (l'île Victoria y compris);
(b) sur la Novaya Ziemlia;
(c) sur la Severnaya Ziemlia (l'île Schmidt et l'île Ouschakov y compris);
(d) sur les îles de Des-Longues.

B. Dans la partie continentale de l'Union Soviétique

(a) sur l'Oural Polaire et l'Oural Prépolaire;

(b) sur la péninsule Taymir;
(c) dans la chaîne de montagnes Kharioulakh (le contrefort septentrional de la chaîne de montagens de Verkhoïansk);
(d) sur le pays d'amont de la rivière Indiguirka;
(e) sur la chaîne de montagnes Koriak (Péninsule Tchoukotsky);
(f) sur la péninsule Kamtchatka;
(g) dans le Caucase;
(h) sur l'Altaï;
(i) sur la chaîne de montagnes Saoursky;
(k) dans les Sayannes;
(l) sur le Thian-Shan (l'Alataou Djoungarsky y compris);
(m) sur le Pamir (l'Alaï y compris).

De cette manière la congélation contemporaine sur le territoire de l'URSS est développée dans 16 régions, dont 4 se trouvent sur les îles des Arctiques Soviétiques et 12 sur la partie continentale de l'Union Soviétique.

Passons à une courte caractéristique des traits principaux de la congélation dans les diverses régions du territoire de l'URSS.

A. Sur les iles des Arctiques Sovietiques

(a) *Congélation de l'Archipel François-Joseph*

L'archipel François-Joseph est couvert presque pour 90 pour cent par des glaciers; ces glaciers forment ici le principal paysage qui y domine. Les 10 pour cent de terre ferme exempts de glace sont représentés principalement par des champs de Zandre et par des affleurements de roches basiques — des basaltes. La superficie générale des glaciers est près de 16000 km carrés.

La congélation de l'archipel François-Joseph répond au type de recouvrement.

Les îles de l'Archipel sont dans la plupart des cas couvertes par des coupoles complètes de glace; leurs parties de bordure s'abaissent vers la mer par des larges traînes et par des torrents de glace isolés. Ces derniers forment souvent des massifs et des langues en éventail de glace flottante. Certaines îles, à relief relativement démembré, portent des recouvrements de glace à surface accidentée qui reflètent le relief basique d'une manière affaiblie. On peut observer ici très souvent plusieurs coupoles divisées par des surfaces exemples de glace ou bien qui sont couvertes comme par des glaciers de vallées.

La limite des neiges éternelles sur les glaciers de l'Archipel François-Joseph se trouve à une petite altitude et elle descend par place jusqu'au niveau de la mer. La puissance de la glace des glaciers varie de 100 à 500 m.

En été les parties supérieures de certaines coupoles glaciaires plus basses et plus petites que les autres sont exemptes de glace et de névé. Ces glaciers sont entièrement formés par la glace des gletschers. Les coupoles plus hautes et plus grandes, situées dans les parties centrales, sont recouvertes par des masses suffisamment puissantes de névé. L'absence mentionnée du névé sur certains glaciers indique que durant la période tiède le restant de l'accumulation annuelle se transforme ici entièrement vers la fin de la période d'ablation en une nouvelle couche de glace « apposée ». Ceci se fait grâce à une quantité suffisante d'eaux dégelées et au petit volume d'accumulation de la neige.

La formation de la glâce sur les glaciers a lieu selon le type d'infiltration congélationnée. Sur les glaciers qui possèdent des masses de névé l'on trouve divers types de formation de la glace : dans les parties inférieures, c'est celui d'infiltration congélationnée, remplacé ensuite par le type froid d'infiltration qui se change dans les parties centrales en type de recristallisation.

Les masses de glace des glaciers de l'Archipel François-Joseph ont toujours des températures négatives. C'est seulement durant les mois les plus chauds que l'on observe la fonte de la couche supérieure.

Outre les coupoles de glace — on voit sur l'Archipel François-Joseph des glaciers « apportés » par les vents. Ces derniers sont disposés généralement sous les escarpements, près du littoral. On peut indiquer encore des petits glaciers originaux, comme le sont les « glaciers — amoncellements » situés au sommet des roches et parfois aussi sur les lignes de partages de la glace si ceux-ci sont étroits et à pentes rapides. Leur genèse est liée à une combinaison particulièrement favorable des vents ascendants et des courants d'atmosphère. Par leur volume ce sont les plus petits glaciers de l'Archipel.

(b) *La congélation de la Novaya Zemlia*

La superficie de la congélation sur la Novaya Zemlia égale approximativement 23, 000 km carrés. La congélation est faiblement développé sur l'île méridionale. Rien que sur la partie septentrionale de cette île on trouve une série de petits glaciers de vallée, de et suspendus, un petit glacier de contreforts et des « taches de névés » isolées. La limite des neiges éternelles est ici de l'ordre de 600-700 m.

La masse principale de la congélation sur la Novaya Zemlia est située sur l'île Septentrionale. Dans la partie sud de cette île, en allant du Matotchkin Shar à peu prés jusqu'à la Krestovaya Gouba, la congélation est représenté par des taches isolées de névé, par des gletschers de vallées, ainsi que par des glaciers de Kare et des glaciers suspendus.

Plus vers le Nord entre la Krestovaya Gouba et la Péninsule de l'Amirauté (approximativement 70° de latitude de Nord) est développé un autre type de congélation; celui qui porte le nom de congélation réticulée.

Ici les vallées remplies de glace forment des plateaux de glace à largeur jusqu'à 7-8 km avec une surface faiblement concave, plus haute que la limite des neiges éternelles, et une surface convexe dans la zone d'ablation. La surface des glaciers dans les parties intérieures possède des altitudes de l'ordre de 300-600 m et les massifs montagneux-« nounataki » atteignent près de 600-900 m. Quatre grandes branches glaciaires, qui s'unissent sur la péninsule de l'Amirauté forment un grand glacier de contrefors; par les huit vallées les plus grandes des torrents de glace descendent dans les amonts des fjords vers le niveau de la mer. La limite des neiges éternelles se trouve dans la région de la congélation réticulée en moyenne à une altitude près de 500 m. La superficie générale des glaciers de congélation réticulée comporte près de 3200 m. carrés.

Encore plus vers le nord de la péninsule de l'Admirauté la congélation réticule se transforme en celle de recouvrement. La couverture de glace de la Novaya Zemlia est la plus grande dans les limites de l'Union Soviétique. Sa longueur atteint près de 340 km. Sa largeur maximum est de 70 km, la surface atteint 19,000 m carrés et la puissance moyenne est estimé à 300-400 m. La couverture glaciale se divise par deux dépressions transversales en trois coupoles principales : la Septentrionale, la Principale et la Méridionale. Leurs parties centrales atteignent conformément des altitudes de 800, 1000 et 900 m. La surface des coupoles est coupée par places par des nounatakis.

La bordure occidentale de la couverture de glace se caractérise par une zone de bras glaciaires qui s'écartent de la masse de recouvrement principale et qui sont entre-divisés par de nounatakis de bordure. Ces bras se joignent ensuite sur la plaine littorale en glaciers de l'étage inférieur des contrefors. Sur le littoral occidental vingt cinq extrémités des glaciers se déversent dans la mer.

La bordure orientale de la couverture glaciaire représente une région de grandes

langues de glaces à largeur jusqu'à 20 km, qui prennent source dans les coupoles méridionale et principale et qui sous le nom de glaciers Nansen, Krapotkine et Nordensheld descendent dans la mer. Telles apophyses ne prennent point source dans la coupole septentrionale et ici la partie majeure du littoral est exempt de glace. La limite des neiges éternelles atteint en moyenne une altitude de 300-400 m.

Dans les parties centrales du recouvrement glaciaire se développe le type de récristallisation de formation des glaces, dans les zones plus basses — le type froid d'infiltration et, enfin dans les parties de bordure et sur les bras des glaciers celui d'infiltration congélationnée. Avec cela dans les parties inférieures de la bordure occidentale du recouvrement (en particulier sur les glaciers de l'étage inférieur situés aux flancs des montagnes)on trouve le type d'infiltration « tiède ».

Dans le recouvrement glaciaire de la Novaya Zemlia sont développés les types suivants du régime thermal de la glace : dans les parties centrales supérieures — type polaire et froid; dans les parties de bordure du recouvrement et dans les langues des glaciers du littoral oriental — le type continental. Dans les parties terminales des bras de glaciers et dans les glaciers des contrefors de l'étage inférieur du littoral occidental — le type marin.

Sur la Novaya Zemlia près des escarpenmets de glace et des rochers, on observe un développement étendu de glaciers « apportés » par les vents.

(c) *La congélation de l'Archipel Severnaya Zemlia (îles Schmidt et Ouschakov y compris)*

La congélation de l'Archipel Severnaya Zemlia se rapporte au type de recouvrement à un stade de dégradation. Les parties intérieures des grandes îles de l'Archipel (Komsomoletz, Pioner, Révolution d'Octobre et Bolschevik) sont recouvertes par des boucliers de glace. L'île Komsomoletz est presque entièrement recouverte par un tel bouclier. Sur les autres îles en allant vers le Sud les terrains exempts de glace se rencontrent de plus en plus souvent et les glaciers finissant sur la terre ferme laissent des grands espaces de plaines littorales libres de glace. La congélation de L'île Bolschevik comprend rien que 20 pour cent de la surface totale de l'île. La où les couvertures glaciaires occupent une grande partie des îles. elles forment des langues flottantes en éventail qui s'abaissent vers la mer. Dans les parties septentrionales de l'Archipel ce sont les bordures des couvertures glaciaires mêmes qui touchent sur des grandes étendues directement à la mer.

Les îles Schmidt et Ouschakov sont entièrement recouvertes par des chapeaux de glace, qui s'achèvent presque toujours par des escarpements de glace submergés (dans la mer).

La superficie de la congélation sur la Severnaya Zemlia comporte près de 16,000 m carrés, ce qui égale approximativement 42 pour cent du territoire de cet Archipel. La hauteur des coupoles de glace dans les parties centrales est de l'ordre de 800-900 m. La puissance de la glace égale en moyenne 200-300 m. La limite des neiges éternelles sur la Severnaya Zemlia est sur l'extrémité septentrionale de l'archipel à quelques dizaînes de mètres au-dessus du niveau de la mer, et cette limite atteint vers le sud près de 300 m. Dans les parties centrales des coupoles de glace on trouve les types de formations de glace suivants : celui de recristallisation et celui d'infiltration « froide ». Le type d'infiltration congélationnée est propre au zones de bordure. Quand au régime de température des masses de glace c'est pour les parties centrales le type polaire et froid, et pour les zones de bordure — le type continental. Des glaciers « apportés » par les vents sont développés sur les extrémités escarpées des recouvrements de glace. Sur les parties exemptes de couvertures de glaces et qui ont un relief tranchant et démembré (l'île Révolution d'Octobre) on trouve des glaciers suspendus et de Kare.

(d) *La congélation des îles de Des Longues*

Sur l'Archipel de Des Longues les îles de Bennnet, d'Henriette et de Jeannette sont recouvertes par une superficie de glace ayant près de 67 km carrés. Ceci égale approximativement 52 pour cent de la surface totale de ces îles. Les glaciers représentent sur toutes les îles des coupoles de glace. Ces coupoles sur les îles Bennet et Henriette atteignent respectivement une hauteur de 450 et 315 m, la puissance de la glace est de l'ordre de 120-150 m. Des apophyses isolées, ayant des pales en éventail flottantes, descendent vers la mer. Sur l'île Jeannette la coupole a une forme irrégulière; la puissance de la glace est en moyenne 80 m. Selon le régime de température les parties centrales des coupoles se rapportent au type froid, et les zones de bordure au type continental. Les types de formation des glaces dans les parties centrales sont respectivement du type d'infiltration froide et dans les zones de bordure type d'infiltration congélationnée.

B. Sur la partie Continentale de l'Union Soviétique

(a) *Congélation de l'Oural Polaire et Prépolaire*

La congélation dans le pays d'amont des rivières Oussa et Schoutschaï dans l'Oural Polaire fut découverte en 1952-1953 et est encore étudiée d'une manière insuffisante étudiée sd'une manière insuffisante.

Cette congélation est représentée par une petite quantité de glaciers minimes. la plus grande partie desquels est du type de Kare et par plusieurs glaciers qui occupent les parties les plus hautes des vallées. La longueur des glaciers ne surpasse point 2 km et en moyenne est de l'ordre de 0,5-0,7 m. La puissance de la glace est en moyenne de l'ordre de 50 m. A ce qu'il paraît l'existence actuelle des glaciers de l'Oural Polaire est imposée par une combinaison très favorable des conditions du relief, du régime des vents, de l'exposition et de l'ombrage. Un des traits caractéristiques de la congélation contemporaine de l'Oural Polaire est l'isolement des glaciers. Il est évident que les glaciers de l'Oural Polaire se rapportent par la formation des glaces au type d'infiltration congélationnée et par le régime de température des glaces au type continental. Le nombre total des glaciers de l'Oural Polaire n'est pas encore établi; il comprend, à ce qu'il paraît, pas plus d'une centaine et la superficie générale des recouvrements de glace ne surpasse pas 50 km carrés.

Il existe sur l'Oural et sur sa partie prépolaire une deuxième région de congélation près des montagnes Narody et Sablia, découverte en 1933. Il y a ici 16 glaciers de Kare d'une étendue de 3 km carrés. La longueur du plus grand d'entre eux est près d'un km; leur longueur moyenne n'atteint pas 0.5 km. La puissance moyenne est de l'ordre de 20-30 m. L'existence de ces glaciers s'explique aussi par la combinaison locale de conditions favorables à la congélation. Par le caractère de la formation des glaces ces glaciers se rapportent au type d'infiltration congélationnée et par le régime de températures de la glace au type continental.

(b) *Congélation de la presqu'île de Taïmyr*

Les montagnes Byrranga sur la presqu'île de Taïmyr ne sont encore pas suffisamment étudiées au point de vue de glaciologie. Actuellement l'on sait qu'il existe sur ces montagnes deux petits glaciers qui occupent les pays d'amont des vallées et plusieurs glaciers du type de Kare. La superficie des glaciers connus ne dépasse point 5 km carrés.

(c) *Congélation de la chaîne de montagnes Kharioulakh*

La congélation de la chaîne de montagnes Kharioulkh, qui est un contrefort septentrional des montagnes Verkhoïansk, est représentée par un nombre minime de glaciers de Kare. Leur nombre total est inconnu et est approximativement estimée comme 3 km².

(d) *Congélation dans le pays d'amont de la rivière Indiguirka*

La congélation dans le pays d'amont de la rivière Indiguirka est caractérisée principalement par des petits glaciers de vallée. La longueur moyenne des glaciers est de l'ordre de 4 km, leur puissance — de l'ordre de 100 km. En outre on observe ici un développement suffisamment étendu de glaciers de Kare et suspendus, des taches de neige glacée (de névé) séparées. La masse principale des glaciers se trouve dans deux régions : sur la chaîne de montagnes Sountar-Khagata et sur le massif Bouordakhsky. En outre, dans les trois régions de Muréline, de Koukhtouy et de la montagne Tchen on trouve des accumulations de petits glaciers de Kare. La superficie totale de congélation atteint 279 km carrés; le nombre des glaciers de types divers est 213. A ce qu'il paraît, c'est le type de formation de la glace d'infiltration froide et celui d'infiltration congélationnée qui sont propres à ces glaciers et encore qui concerne le régime de température des masses de glace ce sont les types froid et continental.

(e) *Congélation de la chaîne de montagnes Koriak* (péninsule Tchoukotsky)

La congélation de la chaîne de montagnes Koriak est représentée principalemens par des glaciers de Kare. L'on y rencontre aussi des glaciers suspendus, des tachet séparées de neige gelée (de névé) et même un certain nombre de petits glaciers de vallée. Leur quantité totale atteint 41; la superficie de congélation est près de 50 km², la puissance moyenne de la glace est de 100 à 50 m. La limite des neiges éternelles égale approximativement 100 m. A ce qu'il paraît, deux types de formations de la glace sont propres à ces glaciers : celui d'infiltration congélationnée et celui d'infiltration « tiède »; il y a de même deux types de régime de température de la glace : type continental et type marin.

(f) *Congélation de la péninsule Kamchatsky*

La congélation de la péninsule Kamchatsky est de diverses formes. Beaucoup de sommets des montagnes et des volcans sont recouverts par des chapeaux en glaces, sur d'autres l'on trouve un développement du type étoilé de congélation. Dans les cratères de certains volcans on observe des glaciers de caldéras; le pays d'amont de certaines vallées est occupé par des gletschers de vallées. Néanmoins c'est principalement le type des glaciers suspendus et de Kare qui est développé ici; c'est glaciers se disposent généralement auprès des crêtes des montagnes et dans les parties supérieures des volcans.

La superficie totale de la congélation est ici près de 866 m², cela fait rien que 0,3 pour cent de la surface de la péninsule.

Sur cette étendue près de 484 km carrés se concentrent dans plusieurs « noeuds » de congélation sur la chaîne de montagne Médiane; près de 34 km carrés sur la chaîne de montagne Orientale; 86 km carrés sur le massif de la presqu'île Kronotsky et 262 km sur les sommets de volcans isolés. La limite des neiges éternelles varie approximativement de 1000 à 1800 m et s'élève en allant du Nord au Sud. On note une position insolite de cette limite sur les volcans; ici son altitude varie de 1900

à 3700 m. A ce qu'il paraît les glaciers de la péninsule Kamchatsky possèdent le type d'infiltration « tiède » pour la formation des glaces et le type marin pour le régime de température des glaces.

(g) *Congélation du Caucase*

La congélation du Caucase est typique pour les montagnes. Cette congélation est proche à la congélation alpine et elle est représentée par des types différents de glaciers, à commencer par les glaciers complexes des vallées jusqu'aux petits glaciers de Kare et suspendus. En outre, il y a au Caucase encore des glaciers du type de buttes volcaniques (sur l'Elbrouz et le Kazbegui). Ces sommets sont recouverts par des chapeaux de glace, dont se séparent des apophyses isolées qui descendent.

La masse principale de la congélation est constituée par des glaciers de vallées; entre ceux-ci on remarque des glaciers de vallées originaux en forme de « T » qui sont distribués particulièrement sur les pentes septentrionales du Caucase Central.

La masse essentielle des glaciers au Caucase (près de 70 pour cent) est disposée sur le versant septentrional et rien que près de 30 pour cent se trouve sur celui du Sud. Les noeuds principaux de congélation sont : les intersections de la chaîne de montagnes qui fait partage des eaux par ses contrefors les plus élevés; le groupe d'Elbrouz, la région du col Mestioisky, le groupe de Schkhari Dykh-Taou, les groupes Tikhtenguensky, Labodsky, Sougansky Mamissonsky, Kazbeguesky.

La superficie totale des glaciers du Caucase surpasse 2200 km carrés et le nombre des glaciers de tous types est plus de 1800. A ce qu'il paraît, après la précision de ces chiffres, ils augmenteront.

Sur le Petit Caucase la congélation est développée faiblement et on la rencontre seulement dans quatres points; sur l'Alagueuse et l'Ararate, ainsi que sur ceux sommets de la chaîne de montagnes Zanzézour. La superficie totale de la congélation sur le Petit Caucase est de l'ordre de plusieurs dizaines de km², avec un chiffre approximativement proche du nombre des glaciers.

La limite des neiges éternelles atteint dans le Caucase Occidental les points entre 2700-2900 km; dans le Caucase Central — 3100-3200 m, dans le Caucase Oriental — 3500-3900 m; et dans le Petit Caucase 3700-4250 m. Deux types de formation des glaces sont propres aux glaciers du Caucase. Pour les glaciers du versant sud et pour une partie de glaciers du versant septentrional situés dans la partie occidentale et centrale — c'est le type d'infiltration « tiède » et le type marin pour le régime de température. Pour la plupart des glaciers de la partie orientale du versant septentrional — c'est le type d'infiltration congélationnée et respectivement le type continental du régime de température des glaces.

(h) *Congélation de l'Altaï*

La congélation de l'Altaï se caractérise par des « petites formes »; elle est représentée en général par des glaciers de Kare et suspendus. Outre ces types on rencontre ici des glaciers de vallées, des glaciers de vallées suspendues et des glaciers de sommets planes. La superficie totale de la congélation de l'Altaï (dans les limites de l'URSS) égale 596 km² et le nombre des glaciers de différents types — 754. La puissance de la glace est en moyenne de 60 à 100 m. La congélation se concentre dans une série de régions, dont les principales sont : le noeud de montagnes Tabyn-Bogdo-Ola et les chaînes de montagnes de l'Altaï Méridional, Katounsky du Tchouysk Méridional, du Tchouysk-Septentrional et celui de Kara-Alakhinski. La limite des neiges éternelles varie de 2000 à 3300 m. On peut supposer que le type dominant de la formation des glaces est celui d'infiltration congélationnée et que le type principal du régime de température est continental.

(i) *Congélation de la chaîne de montagnes Saourski*

La congélation de la chaîne de montagnes Saourski se trouve sur la partie orientale de la crête, qui porte le nom de Mouss-Taou. (La partie ouest de cette crête est disposée dans les limitees de l'URSS). Ici la congélation occupe près de 60 km², dont 35 km² sont dans les limites de l'Union Soviétique. On connaît sur le territoire de l'URSS un total de 10 glaciers, dont 3 sont des glaciers de vallées, 5 de Kare, et 2 suspendus. La limite des neiges éternelles varie de 3300 à 3350 m.

(j) *Congélation des Sayannes*

La congélation des Sayannes se concentre dans deux régions : à Kizir-Kazyrsk et sur le massif Mounkaou-Sardyk; elle est représentée principalement par des taches de neige gelée (du névé) et par un nombre médiocre de glaciers suspendus et de petits glaciers de vallées. Le nombre des glaciers atteint 9 et la superficie totale de la congélation ne surpasse point 5 à 6 km². La limite des neiges éternelles sur le massif Mounkou-Sardyk est près de 2940 m.

(k) *Congélation du Thian-Chan (l'Alataou de Djoungarsk y compris)*

La congélation du Thian-Chan, de cet énorme pays montagneux, ne peut être décrite dans un court exposé, Nous sommes forcés de nous limiter seulement à une simple liste des données principales et des chiffres essentiels. Actuellement, on n'a pas encore enrégistré tous les glaciers du Thian-Chan et l'on n'a pas encore terminé les calculs pour les surfaces de congélation. C'est pourquoi les chiffres cités ne peuvent être acceptés comme définitifs.

Tous les types de glaciers propres à la congélation des pays montagneux sont en fin de compte représentés dans le Thian-Chan et avec cela on y trouve tous les variants — les sous-types.

Outre les énormes glaciers complexes de vallées (type de dendrine) auxquels l'on rapporte le plus grand glacier de Thian-Chan — celui qui est nommé Inyltchek (dans le groupe Khan-Tengry), à longueur dépassant 65 km, on y rencontre les sous-types les plus variés de glaciers complexes et simples de vallées, y compris la variété spéciale qui porte le nom de glaciers de « Tourkestan », les glaciers assymmétriques des vallées, les gletschers des cuvettes, les glaciers des vallées suspendues. En outre, on trouve dans le Thian-Chan, sur les parties faiblement inclinées des crêtes et des plateaux les glaciers des sommets planes qui rapellent par leur aspect les recouvrements glaciaires des îles situées dans les Arctiques. On peut citer encore un autre type tel que le type de glacier sur les versants et bien entendu les divers glaciers de Kare et suspendus.

Le type le plus caractéristique pour la masse de congélation contemporaine du Thian-Chan est représenté par les glaciers des vallées de dimension relativement médiocre à longueur de 3 à 5 km.

Les foyers principaux de congélation contemporaine se trouvent dans le noeud montagneux du Khan-Tengry, dans le massif Ak-Schiyriak et dans le massif Talgarsky. La limite des neiges éternelles dans les chaînes de montagnes de bordure est de l'ordre de 3000 à 3600 m et pour les chaînes de montagnes intérieures de 4000 à 4500 m.

La masse principale de congélation répond au type d'infiltration congélationnée de formation de la glace et le type continental du régime des températures de celle-ci. Une exception est représentée par les parties supérieures de la congélation des grands noeuds montagneux et par les élèvements les plus hauts où l'on observe une substitution des types de formation des glaces en allant des parties supérieures aux parties

inférieures. Cela se fait en succession suivante : type de recristallisation, type de recristallisation à infiltration; type froid d'infiltration et ensuite type d'infiltration congélationnée. Une succession du même genre a lieu pour le régime des températures dans la masse des glaces, en commençant par le type polaire sec suivi du type polaire humide et froid et enfin du type continental.

La superficie totale de congélation du Thian-Chan comprend dans les limites de l'URSS approximativement 8700 km carrés, le nombre des glaciers surpasse 1500 (toute une série de glaciers n'étant pas encore enrégistrée).

(l) *Congélation du Pamir et de l'Altaï*

La congélation du Pamir et de l'Altaï, c'est-à-dire du plus grand et du plus haut système de montagnes sur le territoire de l'URSS (de même que la congélation du Thian-Shan) ne peut être décrite dans une courte communication et nous sommes forcés de nous limiter à des données plus générales. Selon la variété des formes de congélation contemporaine, la congélation actuelle du Pamir ne cède en rien au Thian-Chan et elle est représentée en somme par les mêmes types de glaciers. Notons que dans les limites du Pamir se trouve le plus grand glacier de montagne de l'URSS — celui portant le nom de Fedchenko. La longueur de ce dernier surpasse 70 km et sa superficie totale égale 907 km carrés. De même que pour le Thian-Chan, la masse principale de congélation du Pamir est formée par des glaciers de vallées relativement petits, à longueur de 4 à 6 km. En ce qui concerne les types de formation des glaces et du régime des températures de celles-ci, nous trouvons sur le Pamir le même tableau que pour le Thian-Chan.

La surface totale de la congélation du Pamir et de l'Alay égale dans les limites de l'URSS 9276 km², le nombre des glaciers atteint presque 1200 (les petits glaciers à longueur moins de 1,5 km ne sont point inclus dans ce nombre).

Les plus grands centres de congélations contemporaine sur le Pamir et l'Alay sont les chaînes de montagnes Matchinski, ceux de l'Académie des Sciences, du Transalay, du Rouschansko-Bazardarinsky, du Yaz5oulemsky, du Darvazsky, de Pierre le Grand et celui de Zouloumart. La limite minimum des neiges éternelles se trouve sur le Pamir et sur l'Alay à une hauteur de 3400 m (à l'occident) et la limite maximum à près de 5250 m (à l'Orient) au-dessus du niveau de la mer.

La superficie totale de la congélation contemporaine sur le territoire de l'Union Soviétique comprend approximativement 77,146 km², desquels près de 55,067 km carrés sont situés sur les îles des Arctiques et 92,079 km carrés dans les régions montagneuses de la partie continentale du territoire de l'URSS. On compte dans cette dernière plus de 6000 glaciers isolés.

Les données sur la congélation contemporaine du territoire de l'URSS sont exposées selon les sources litéraires nouvelles. Une liste de ces sources se trouve dans le Rapport de l'Union Soviétique de géodésie et de géophysique préparé pour la partie intitulée « Hydrologie scientifique ». Les données indiquées plus haut sont citées de même selon les connaissances propres de l'auteur pour une série de régions de congélation.

V. Travaux Glaciologiques de l'URSS durant l'AGI.

Le but principal des investigations glaciologiques, exécutées durant l'A.G.I. consiste en une étude des processus d'accumulation, de transformation et de débit de la glace sur la surface terrestre en relation au bilan thermal de cette dernière, ainsi qu'en l'étude des glaces et de la circulation d'eau sur notre planète. Les plus importants problèmes qui doivent être étudiés sont : l'action réciproque de la congé-

lation et du climat; l'état actuel, la distribution en espace et la puissance de la congélation, la direction de son évolution présente.

Ce but répond à la tendance contemporaine de la glaciologie dans l'Union Soviétique et fait le développement ultérieur des travaux glaciologiques qui sont exécutés en l'URSS.

En choisissant les points pour l'exécution de travaux glaciologiques sur le territoire de l'URSS durant l'AGI, nous avons été guidés par trois considérations essentielles : (1) de gagner par les investigations les régions principales de congélation; (2) d'inclure dans les investigations les régions à conditions naturelles diverses et (3) d'exécuter ces travaux sur les mêmes objects qui furent examinés durant la Deuxième A.P.I. En outre, l s glaciologues soviétiques avaient un grand intérêt à exécuter des observations dans l'Antarctide, en tant que les investigations de cette région présentent un intérêt scientifique exceptionnel et qu'elles sont tout à fait indispensables pour le développement ultérieur de la glaciologie. Il est évident que l'organisation des investigations glaciologiques répond d'une manière suffisamment complète à ces exigences par l'exécution des travaux indiqueés dans les points suivants :

(1) Sur l'Archipel de François-Joseph, sur l'île Gouker, dans la baie Tikhaya;
(2) dans la Novaya Zemlia sur son île septentrionale dans la région du Port Russe (Rousskaya Gavagne);
(3) sur l'Oural Polaire;
(4) dans les Khibiny;
(5) dans Zagorsk;
(6) au Caucase dans la région de l'Elbrouz;
(7) sur l'Altaï
(8) dans le Thian-Chan sur la chaîne de montagnes Tersky Alataou;
(9) dans le Thian-Chan sur la chaîne de montagnes Zailiysky Alataou;
(10) dans le Pamir sur le glacier Fedchenko;
(11) sur la chaîne de montagnes Sountar-Khayata;
(12) dans l'Antarctide sur la zone littorale et sur la partie intérieure du continent

Dans les régions indiquées sous les Nos. 1, 2, 6, 7, 8, 9 et 10 des observations glaciologiques furent exécutées durant la 2-ème A.P.I. Les congélations sur l'Oural Polaire et sur la chaîne de montagnes Sountar-Khayata sont peu étudiées et elles se caractérisent par des particularités spécifiques, c'est pourquoi elles présentent un grand intérêt à être étudiées.

Les observations dans les Khibiny, c'est-à-dire dans une région qui se trouve sur le seuil d'un développement possible de congélation représente aussi un intérêt de même grandeur. Enfin, l'on propose d'exécuter à Zagorsk des investigations expérimentales et comparatives.

Avant le commencement des investigations glaciologiques on exécuta un immense travail en vue d'élaborer des indications méthodiques pour tous les aspects des observations glaciologiques. Nous avons aspiré de cette manière à assurer la possibilité d'une plus grande uniformité des résultats des travaux, une telle uniformité est nécessaire pour une comparaison du matériel obtenu, ainsi que pour l'exécution des travaux à un niveau scientifique plus élevé.

Nous supposons que le choix des régions à observer et l'exéctution de toutes les investigations selon un programme unique permets d'espérer que l'on obtiendra une grande quantité de matériel et de faits à base desquels l'on pourra faire avancer d'une manière importante nos connaissances des lois naturelles qui sont propres aux phénomènes de glaciologie.

Actuellement, on exécute déja, dans toutes les régions glaciaires énumérées plus haut, des observations glaciologiques selon le programme de l'A.G.I. Ces observations vont se prolonger sans interruption durant toute l'A.G.I.

Au moment présent il est certainement prématuré de parler des résultats de ces travaux.

Comme on le sait, les travaux dans les Antarctiques ont commencé avant les observations dans d'autres régions et l'on connaît déjà certains résultats préalables ayant un intérêt spécial qui porte sur les particularités de la structure du recouvrement glacial de l'Antarctide, sur le processus de la formuation des glaces, sur le régime des températures, sur la puissance des glaces, etc. J'ai fait à ce propos une communication à la Quatrième Conférence Antarctique à Paris.

Je termine ainsi mon exposé concernant les inverstigations glaciologiques dans l'Union Soviétique. Nous avons supposé qu'une telle communication serait utile en tant que les glaciologues soviétiques prennent pour la première fois la parole dans l'Assemblée de l'Union Internationale de Géodésie et de Géophysique.

G. A. Avsiuk, professeur.

BEITRÄGE ZUR ERFORSCHUNG SUBGLAZIALER TALFORMEN UND DER IN IHNEN LIEGENDEN ABLAGERUNGEN

O. FOERTSCH und H. VIDAL. (*)

Abstract

A team of the Institute for Applied Geophysics of the University of Munich has taken seismic refraktion measurements on several glaciers of the Austrian Alps in order to study the elastic behaviour of the ice, ground moraine and bedrock, to determine the thickness and volume of the ice and ground moraine and to explore the morphology of the bedrock of the glaciered valleys.

Von jeher war die Glaziologie an der Erkundung der Gestalt des subglazialen Untergrundes und der Ermittlung der in Gletschern gespeicherten Eisvolumina interessiert. Man hat versucht, durch Bohrungen und durch den Vortrieb von Stollen Einblicke in die Eisverhältnisse zu gewinnen. Da wegen der Kostspieligkeit solcher Arbeiten nur wenige derartige Aufschlüße angelegt werden können, hat man frühzeitig nach anderen Möglichkeiten zur Ablotung von Gletschern Ausschau gehalten. In erster Linie ist hierbei an die Heranziehung verschiedener geophysikalischer Aufschlußverfahren zu denken. Es hat sich dabei gezeigt, daß für eine ausgedehnte räumliche Erfassung der Untergrundsverhältnisse unter den Gletschern die Refraktions- und Reflexionsverfahren der angewandten Seismik am geeignetsten sind, wobei die nötigen Schallwellen durch kleine Sprengungen von einigen Kilogramm Sprengstoff erzeugt werden.

Auf Anregung von Prof. R. Finsterwalder hat das Institut für angewandte Geophysik der Universität München 1953 die seismische Untersuchung ostalpiner Gletscher in sein Arbeitsprogramm aufgenommen. Seit dieser Zeit wurden vermessen :

1953 Teile des Gepatsch- und Kesselwandferners (Oetztaler Alpen)
1954 Hintereisferner (Oetztaler Alpen)
1955 Brandnerferner (Rätikon)
1956 Großer Gurglerferner (Oetztaler Alpen)

Der Brandnerferner ist kein eigentlicher Gletscher sondern eher als Firn-bzw. Gletscherfleck zu bezeichnen. Alle anderen sind typische Talgletscher mit Firnfeld und ausgeprägter Zunge.

Die Durchführung der Messungen wurde durch die finanzielle Unterstützung der Deutschen Forschungsgemeinschaft und der Vorarlberger Illwerke ermöglicht, denen auch an dieser Stelle aufrichtig gedankt sei.

Die Ausbreitung der Schallwellen im Eis und in seiner Unterlage wurde mit der « Geophonapparatur » des Institutes beobachtet, die im Laufe der Jahre ständig für die Arbeiten auf Gletschern verbessert worden ist. Die Erschütterungsaufnehmer, Geophone, wandeln die kleinen Bewegungen, in die sie durch die ankommenden Schallwellen versetzt werden, durch Druckänderungen auf Mikrophonkapseln in Schwankungen des Mikrophonstromes um. Diese werden an der zentralen Regi-

(*) Anschriften der Verfasser : Dr. O. Förtsch, Institut für angewandte Geophysik der Universität, München 2, Richard Wagnerstr. 10 .
Dr. H. Vidal, Bayerische Landesanstalt für Moorwirtschaft und Landkultur, München 23, Leopoldstr. 5.

strierstation mittels Transformatoren von Ruhestrom getrennt und mittels kleiner Galvanometer auf Registrierpapier aufgezeichnet. Der Umfang der Apparatur ist im Laufe der Jahre von 6 auf 12 Geophone erweitert worden. Neben den von den Geophonen aufgeschriebenen Erschütterungen wird weiterhin auf den Registrierungen noch der Moment der Sprengung in der üblichen Weise durch Zerreißen eines stromdurchfloßenen Drahtes bei der Detonation vermerkt. Bei den Beobachtungen werden die Geophone jeweils auf Meßstrecken, Profilen, verteilt, deren Längen sich nach den örtlichen Gegebenheiten richten. Auf den Gletschern variierten sie zwischen 400 und 1000 m. Am Anfang und meist auch am Ende eines jeden Profils sind Sprengungen abgetan worden, deren Ladungen zwischen 1 und 5 kg lagen. Sie wurden in Bohrlöchern bis 2 m Tiefe oder auch in kleinen Vertiefungen der Eisoberfläche gezündet.

Die durch die Sprengungen erzeugten Erschütterungen breiten sich teils als longitudinale und transversale Wellen im Eis aus, in nahen Entfernungen vom Sprengort werden sie an der Unterlage des Eises reflektiert und kommen als Echos zur Oberfläche zurück, teils dringen sie in den tieferen Untergrund ein, laufen an den Grenzflächen desselben entlang und werden dabei zur Eisoberfläche zurückgestrahlt. Aus den von den Geophonen vermerkten Ankunftszeiten der verschiedenen Wellen läßt sich der Verlauf der Grenzflächen im Untergrund berechnen. Da die Grenzflächen im allgemeinen nicht eben sind, dürfen für die Berechnungen nicht die in der Seismik sonst gebräuchlichen, einfachen Beziehungen für solche Flächen herangezogen werden, sondern man muß aus jeder beobachteten Ankunftszeit individuell die entsprechenden Tiefen berechnen, wobei die Beobachtungen bei den Sprengungen an den Endpunkten der Profile und die Tiefenberechnungen für die Schnittpunkte zweier Profile als wichtige Kontrollen dienen. Um einen Einblick zu geben, wie die Profile bei einer seismischen Vermessung eines Gletschers angelegt werden, ist in Abb. 1 ein Plan des Gurglerferners mit den eingezeichneten Beobachtungslinien wiedergegeben.

Abb. 1

Beobachtete Geschwindigkeitswerte. Aus den Registrierungen lassen sich als erstes die Ausbreitungsgeschwindigkeiten der Longitudinal- und Transversalwellen im Eiskörper und der Longitudinalwellen in der unter dem Eis folgenden Grundmoräne sowie im Felsuntergrund bestimmen.

a. Die Longitudinalgeschwindigkeit im Eis.

Die ermittelten Werte für diese Geschwindigkeit schwanken von Beobachtung zu Beobachtung sehr stark. Dies hängt nicht nur mit der sich von Ort zu Ort ändernden Konsistenz des Eises, sondern vermutlich noch mehr mit der Schichtung desselben zusammen, da sich auch auf dem gleichen Profil bei entgegengesetzter Ausbreitungsrichtung der Wellen verschiedene Ausbreitungsgeschwindigkeiten ergeben. Für den Hintereisferner liegen die gemessenen Werte zwischen 3500 und 3850 m/s und steigen vereinzelt sogar auf 4000 m/s an. Niedrigere Werte ergaben sich auf der Zunge des Gurglerferners. Hier streuen sie zwischen 3425 und 3600 m/s. Noch kleinere Geschwindigkeiten bis 3000 m/s herab wurden auf dem Kesselwand- und Gepatschferner in Höhen über 3100 m über NN beobachtet. Mitten im großen Firngebiet des Gepatschferners pflanzten sich die Wellen nur mit 1600 m/s fort. Das Eis des Brandnerferners besitzt eine mittlere Eisgeschwindigkeit von etwa 2900 m/s. Wenn auch zweifellos eine Tendenz dahin zu erkennen ist, daß die Schallgeschwindigkeit beim Übergang vom Eis zum Firn auf den Ostalpengletschern abnimmt, so ist es doch schwer, irgendwelche Gesetzmäßigkeiten zu erkennen.

Bei der Berechnung der Mächtigkeiten der Schichten des Untergrundes hat sich ergeben, daß eine gegenseitige Übereinstimmung der Zahlenwerte nur dann zu erzielen war, wenn in die Rechnung eine einheitliche Geschwindigkeit für das gesamte Eispaket eingesetzt wurde. Dies deutet darauf hin, daß die beobachteten Geschwindigkeitswerte nur oberflächennahen, relativ dünnen Eisschichten zugeordnet werden dürfen, während das Eis in seiner Gesamtmächtigkeit doch eine gleichbleibende mittlere Geschwindigkeit für Schallwellen besitzt. Allgemein wurde den Berechnungen der Wert 3600 m/s zugrunde gelegt, da dieser als der wahrscheinlichste für Eis der Temperatur 0° C anzusehen ist. Da dieser Mittelwert aber nicht unmittelbar aus den Beobachtungen abgeleitet werden kann, ist darin eine gewiße systematische Unsicherheit der Mächtigkeitsberechnungen für die verschiedenen Schichten des Untergrundes begründet.

b. Die Transversalgeschwindigkeit im Eis.

Für die Transversalgeschwindigkeiten im Eis gilt hinsichtlich der Schwankungen der Beobachtungswerte das gleiche wie für die Longitudinalgeschwindigkeiten, falls derartige Wellen überhaupt angeregt werden. Im Firngebiet des Kesselwand- und Gepatschferners konnten in den Registrierungen keine Transversalwellen erkannt werden. Das war auch auf dem Brandnerferner der Fall. Dagegen sind im Zungeneis des Hintereisferners und des Gurglerferners Transversalwellen mit großer Amplitude aufgetreten.

Auf dem Hintereisferner schwanken die beobachteten Transversalgeschwindigkeiten zwischen 1600 und 1925 m/s, während sie auf dem Gurglerferner zwischen 1300 und 1920 m/s streuen. Als Mittelwert ergibt sich 1730 bzw. 1630 m/s. Die Berechnung der Poissonschen Zahl aus den Mittelwerten der Longitudinal- und Transversalgeschwindigkeit liefert für beide Gletscher den Wert 0,36, wie er sich auch bei früheren Beobachtungen auf anderen Gletschern ergeben hat.

Hier darf vielleicht noch auf eine seismologisch interessante Erscheinung hingewiesen werden. Bei anderen Refraktionsbeobachtungen (REICH 1952) wurde festgestellt, daß der Welle im oberen Medium oder auch der Refraktionswelle in bestimmten zeitlichen Abständen Wellen folgen, die sich mit der gleichen Geschwindigkeit wie diese, nämlich mit der Geschwindigkeit des betreffenden Mediums, ausbreiten. Auf dem Hintereisferner und auch auf dem Gurglerferner konnten auf je einem Profil solche Wellen, die REICH (1952) reflektierte Refraktionsimpulse genannt hat, registriert werden, die mit der Geschwindigkeit der Transversalwelle im Eis dieser mehrmals nacheinander in bestimmten zeitlichen Abständen folgen. Für die Deutung dieser Erscheinung dürfte gerade das Auftreten der transversalen «reflektierten Wellen» von großer Bedeutung sein. Allerdings kann heute eine befriedigende Erklärung dieser Wellen noch nicht gegeben werden.

c. Die Longitudinalgeschwindigkeit in der Grundmoraene.

Die Beobachtung der Refraktionswellen gibt uns Aufschluß über die Beschaffenheit der Unterlage des Eises. Unter großen Teilen der untersuchten Gletscher wurden für diese Longitudinalgeschwindigkeiten zwischen 3700 und 5200 m/s mit Schwerpunkten zwischen 4000 und 4500 m/s beobachtet. Es zeichnen sich gewiße Zusammenhänge der Geschwindigkeiten mit der Mächtigkeit und der Dynamik des Eises ab. Man hat aber auch im Vorfeld von Gletschern unter den Obermoränen derartige Geschwindigkeiten gemessen. Zweifellos können solche Geschwindigkeiten nur Grundmoräne zugeordnet werden, wobei es erstaunlich ist, daß in derartigen, jungen Ablagerungen so hohe Werte auftreten können. Verständlich würden sie durch die Annahme werden, daß sich die Grundmoräne im gefrorenen Zustand befindet. Es kann jedoch auch sein, daß die hohe Ausbreitungsgeschwindigkeit durch eine besonders dichte, porenvolumenarme Sedimentation des Materials und durch eine durch Bewegung und Druck der Eislast verursachte Verdichtung (Dynamodiagenese) bedingt wird. Nicht nur unter den Zungen des Hintereis- und Gurglerferners, sondern auch im Firnbereich des Kesselwand- und Gepatschferners konnte Grundmoräne festgestellt werden. Unter dem Gurglerferner ließ sich sogar ihre Mächtigkeit errechnen, was beim Hintereisferner nur im Vorfeld möglich war. Die Ausbreitungsgeschwindigkeiten in der Obermoräne sind kleiner als 3000 m/s.

d. Die Longitudinalgeschwindigkeit im Felsuntergrund.

Auf einigen Profilen des Gepatsch- und des Hintereisferners wurden für die Unterlage des Eises Geschwindigkeiten über 5400 m/s gemessen. Hier steht ohne Zweifel unter dem Eise der gewachsene Fels an. Auf dem Gurglerferner konnte für jedes Profil die Felsgeschwindigkeit bestimmt werden. Die Werte schwanken zwischen 4910 und 6050 m/s. Der Mittelwert liegt bei etwa 5400 m/s.

Die Unterlage der Gletscher und die glazialerosive Ausformung des Felsuntergrundes. Von den seismisch untersuchten Gletschern konnte nur am Gurglerferner der Verlauf der Felssohle des Taltroges ermittelt werden, während bei den anderen nur die unmittelbare Unterlage des Gletschers, d. h. im wesentlichen der Verlauf der Grundmoränenoberfläche, erfaßt werden konnte. Für den Brandner-, Hintereis- und Gurglerferner wurden für die Grundmoränenoberfläche, für den Gurglerferner auch für die Felsoberfläche Höhenschichtlinienpläne gezeichnet und daraus wiederum Berechnungen über das Eis- bzw. Grundmoränenvolumen angestellt und Gletschermodelle angefertigt, die eine gute Vorstellung von der Ausformung des Glestcherbettes vermitteln.

Das Bett des *Kesselwandferners* bildet im Bereich des Firnfeldes eine flache und schwach reliefierte Mulde mit maximalen Eismächtigkeiten von 120 m und einer durchschnittlichen Eisdicke von 80 m. Mit Annäherung an den oberen Zungenbereich verengt sich das Gletscherbett und muldet sich stärker ein, wobei Eisdicken bis 200 m erreicht werden.

Die drei auf dem Firnfeld des Gepatschferners seismisch vermessenen Profile zeigen ebenfalls nur geringe Reliefunterschiede des Gletscherbettes. Die maximalen Eisdicken erreichen hier 150 m, die durchschnittliche Tiefe liegt bei 100 m. An der Eisscheide zwischen Kesselwand- und Gepatschferner, am Kesselwandjoch, liegen noch 60 m Eis.

Die wesentlich umfangreicheren seismischen Messungen auf dem *Hintereisferner* ergaben ein ziemlich genaues Bild von der Form des Eiskörpers und seiner Unterlage (Abb. 2). Die in Abb. 2 unten wiedergegebenen Querschnitte zeigen, daß das Eis

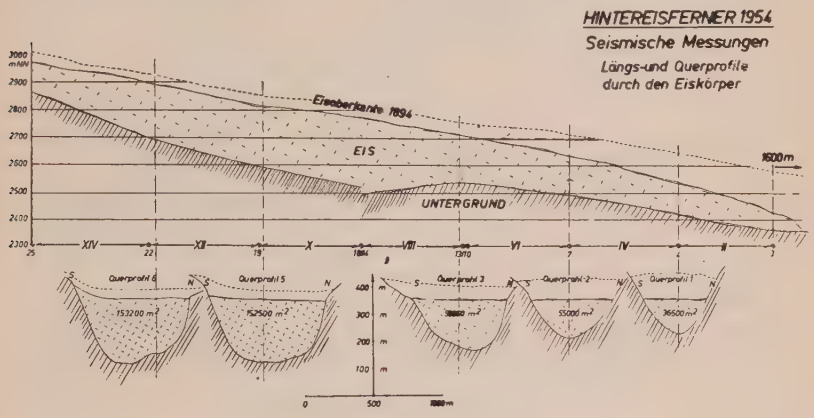
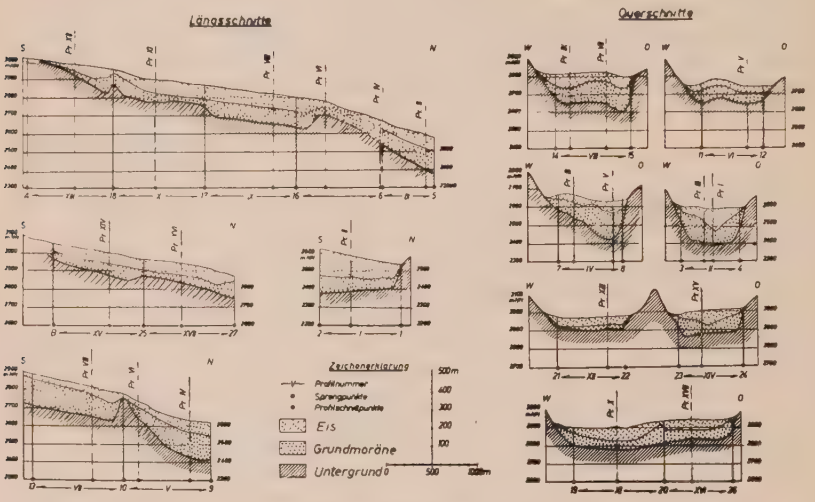


Abb. 2

im Zungenbereich in einem U-förmig ausgebildeten Bett liegt, welches sich mit Annäherung an die Firngrenze muldenförmig weitet. Ferner läßt sich eine deutliche Asymmetrie des Talquerschnittes feststellen, die durch verschieden starke Erosionswirkung des Eisstromes an den prall-bzw. gleithangseitigen Kurvenpartien bedingt ist. Das Längsprofil (Abb. 2, oben) gibt einen Einblick in die Gefällsverhältnisse des Gletscherbettes im Zungenbereich. Das Sohlengefälle beträgt im unteren Teil etwa 9 %, verflacht sich dann noch etwas gletscheraufwärts, um etwa in Zungenmitte eine Schwelle — vermutlich Abzeichnung einer Trogschwelle im Felsuntergrund — zu überschreiten, dann rückläufig zu werden und 600 m gletscheraufwärts einen neuen Tiefpunkt zu erreichen. Dieser zeichnet vermutlich das Zentrum eines Trogbeckens im Felsuntergrund ab, das hier im Brennpunkt glazialer Erosionswirkung liegt. Der aus dem Firnfeld steil herabkommende Eisstrom trifft auf eine Flachstrecke. Gleichzeitig erfährt er hier einen Massenzuwachs durch den über einen steilen Eisbruch einmündenden Nebengletscher, den Laugtauferer-Joch-Ferner, wodurch die Erosion noch verstärkt wird. Wir haben es also mit einem sog. Mündungsbecken (R. v. Klebelsberg 1948) zu tun. Von hier aus steigt dann die Gletschersohle mit einer gleichmäßigen Steigung von 15 bis 20 % zum Firnbereich an. Die Eismächtigkeiten schwanken zwischen 130 und 290 m.

Dagegen sind die Eisdicken des *Brandnerferners* sehr klein. Sie erreichen nur 20 bis 25 m.

Am *Gurglerferner* (Abb. 1) war es den Verfassern durch das Zusammentreffen mehrerer günstiger Umstände, nämlich relativ geringe Eis- und Grundmoränenmächtigkeiten und lange Beobachtungsprofile, gelungen, die Grenzflächen Eis-Grundmoräne bzw. Grundmoräne-Felsuntergrund getrennt ermitteln zu können. Dadurch wurde es möglich, ein detailliertes Bild von den Volumenverhältnissen der Eis- bzw. Grundmoränenfüllung des Gletschertroges und seiner Morphologie zu entwerfen. Die Längs- und Querschnitte der Abb. 3 veranschaulichen die seismischen Meßergebnisse.



Seismische Messungen auf dem Gurgler-Ferner 1958

Abb. 3

Der Gurglerferner wird im Firngebiet durch den Mitterkamm (Abb. 1) in zwei Eisströme geteilt. Die seismischen Untersuchungen erstreckten sich nur bis zur unteren Grenze des Firns, bis zur Höhenschichtlinie 3000 m über NN. Wegen der starken Absorption elastischer Wellen durch den Firn können sich in ihm solche Wellen nicht ausbreiten. Bereits in diesen Höhen ist unter dem Gletscher Grundmoräne in beachtlicher Dicke vorhanden. Das Bett des schwächeren, östlichen Eisstromes, Profil XV und XVII, verläuft annähernd parallel zur Eisoberfläche. Im Felsuntergrund ist bei Sprengpunkt 25 eine kleine Stufe vorhanden. Ungefähr in gleicher Höhe ist beim westlichen Eisstrom bei Sprengpunkt 18 (Profil XIII, X, IX und III) eine Schwelle aus dem Felsuntergrund herausgearbeitet worden. Das Trogbecken oberhalb und das Fußbecken unterhalb der Schwelle ist deutlich ausgeprägt. In abgeschwächter Form ist das Relief auch in der Oberfläche der Grundmoräne und in etwa auch im Verlauf der Eisoberfläche zu erkennen. Beachtlich ist die große Mächtigkeit der Grundmoräne in den beiden Becken.

In der Mitte der Zunge des Gurglerferners wiederholt sich dieselbe Konfiguration im vergrößerten Maßstab von Sprengpunkt 17 bis Sprengpunkt 5 nochmals. Mit den Parallelprofilen VII und V ist die Schwelle in noch markanterer Weise erfaßt worden. Hier ist der Felsriegel so hoch, daß er die ganze Grundmoräne durchdringt, die ober-und unterhalb desselben unwahrscheinlich stark ausgebildet ist. Die Profile IV und II vermitteln uns Querschnitte durch das Tal des Gurglerferners unterhalb

der großen Schwelle. Unmittelbar unterhalb derselben (Profil IV) ist eine starke Asymmetrie vorhanden. Der tiefe Einschnitt in die Grundmoräne darf wohl auf die Erosion des unterirdisch abfliessenden Schmelzwassers zurückgeführt werden. Das Zungenende des Gletschers liegt heute in einer schmalen Klamm. Auf kurze Erstreckung fällt das Eis 200 m tief zum Gletschertor hinab. Die seismischen Messungen endeten am Beginn des Steilabbruches.

Die geschilderten Ergebnisse der seismischen Untersuchungen geben einen guten Einblick in die Morphologie des Gletscherbettes und des Felsuntergrundes und vermitteln wichtige Erkenntnisse über die Dynamik des Eises. Im Gegensatz zum Wasser gräbt sich Eis vor einem Hindernis in sein Bett ein und arbeitet so ein Trogbecken heraus. Beim Herabfließen über eine subglaziale Felsschwelle (Riegel) erhöht sich die Fließgeschwindigkeit des Eises. Beim Übergang in ein flacheres Sohlengefälle tritt eine Kraftkomponente in Aktion, die im Untergrund ein zweites Becken, das sog. Fußbecken, anlegt. Die beobachteten starken Übertiefungen vor und hinter den Riegeln vermitteln uns einen Maßstab für die Beträge, die die Glazialerosion erreichen kann (R. v. Klebelsberg 1948/I,S.362).

Es wurde schon erwähnt, daß unter dem Eis eine erstaunlich mächtige Grundmoräne vorgefunden wurde, und daß sich in ihrer Oberfläche in abgeschwächtem Maße das Relief des Felsuntergrundes wiederspiegelt. Dies läßt darauf schließen, daß sie nicht gleichmäßig abgelagert wird, sondern nach einer anderen, durch die Dynamik des Gletschereises festgelegten Gesetzmäßigkeit. Das Vorhandensein der Felsschwellen beweist, daß der Felsuntergrund des Gurglerferners sicher durch Eis überformt worden ist. Dies muß zu einer Zeit geschehen sein, als im Gurglertal mehr Eis lag als heute. Die Ausbildung der Grundmoräne deutet weiter darauf hin, daß mit ab- oder zunehmender Eismächtigkeit das Gletscherbett den neuen dynamischen Verhältnissen nicht mehr angepaßt ist. Der Gletscher ist bestrebt sich eine bessere Anpassung zu schaffen. Bei einem Rückzug des Eises wird an seiner Unterkante Grundmoräne sedimentiert. Im Falle eines Vorstoßes schiebt das Eis eine entsprechende Schicht Grundmoräne vor seinem Zungenende her und türmt sie dort als Endmoränenwall (Stauchmoräne) auf. Zu der Ab- bzw. Zunahme eines Gletschers, die man durch ein Absinken bzw. Anheben seiner Oberfläche feststellen kann, kommt noch die, die durch das entsprechende Anheben bzw. Absinken der Eisunterkante bedingt ist. Die Grundmoräne ist wahrscheinlich wesentlich stärker in die Dynamik der Gletscher miteinbezogen, als man das bisher vermutet hat. Erwähnt sei in diesem Zusammenhang noch, daß auch unter dem Eispanzer Grönlands Grundmoräne mit einer Ausbreitungsgeschwindigkeit von 4800 m/s festgestellt worden ist (Joset, Holtzscherer 1953 und Holtzscherer 1954).

Am Zusammenfluß der beiden Eisströme unterhalb des Mitterkammes bildet sich wie üblich eine Mittelmoräne aus. Die seismischen Ergebnisse bestätigen die aus anderen Beobachtungen (R. v. Klebelsberg 1948) abgeleitete Feststellung, daß der wesentliche Bestandteil einer Mittelmoräne ein subglazialer Moränenwall ist. Die seismischen Messungen bestätigen weiterhin, daß sich dieser Wall gletscherabwärts mehr und mehr verflacht. Auf dem Gurglerferner endet der Wall an der großen Schwelle im mittleren Teil des Gletschers. Da wo er auftrifft, wird vom Eis mehr Moränenmaterial über den Riegel geschleppt und unterhalb auf einem Schuttkegel abgelagert, der sich nicht als Wall fortsetzt. Vom Schuttkegel ist an der Eisoberfläche nichts zu erkennen. Ein Teil des mitgeführten Materials tritt an dieser Stelle durch die relativ dünne Eisdecke auf Scherflächen an die Oberfläche und wird als « Scheinmittelmoräne » gletscherabwärts transportiert.

Die Eis- und Grundmoränenvolumina der Gletscher. Wie schon erwähnt, gestatten die seismisch bestimmten Eis- und Grundmoränenmächtigkeiten auch eine Volumenberechnung dieser Körper. In der Tabelle sind die Volumina für die seismisch unter-

suchten Teile der Gletscher zusammengestellt. In den Spalten « Gesamtvolumen » sind zu den vorstehenden Werten noch die geschätzten Volumen der nicht vermessenen Gebiete addiert.

TABELLE

Gletscher	Berechnetes Volumen Eis	entsprechende Wassermenge	Gesamtvolumen Eis	entsprechende Wassermenge
Kesselwandferner	0,2 km³	0,18 km³	0,3 km³	0,27 km³
Hintereisferner	0,44 km³	0,40 km³	0,84 km³	0,68 km³
Brandnerferner	0,027 km³	0,022 km³	0,027 km³	0,022 km³
Gurglerferner	0,24 km³	0,21 km³	0,32 km³	0,26 km³
		Grundmoränenvolumen		
Gurglerferner	0,40 km³		0,47 km³	

Im Literaturverzeichnis sind auch die Veröffentlichungen der Verfasser aufgeführt, in denen über die oben referierten Arbeiten eingehender berichtet wird.

SUMMARY

A team of the Institute for Applied Geophysics of the University of Munich has taken seismic refraction measurements on several glaciers of the Austrian Alps in order to study the elastic behaviour of the ice, ground moraine and bedrock, to determine the thickness and volume of the ice and ground moraine and to explore the morphology of the bedrock of the glaciered valleys. For this purpose measurements were taken along transverse and longitudinal sections.

Longitudinal velocities in the ice were found to vary from 3500 to 3850 m/s on the tongue of the Hintereisferner, from 3425 to 3600 m/s on the tongue of the Gurglerferner, from 1600 to 3000 m/s on the snow-fields of the Gepatschferner and Kesselwandferner and with 2900 m/s on the snowfield of the Brandnerferner.

The *transverse velocities* differ between 1600 and 1925 m/s on the Hintereisferner and 1300 to 1920 m/s on the Gurglerferner with a mean value of 1730 and 1630 m/s respectively. The POISSON number is 0.36.

The true *longitudinal velocities in the ground moraine* under the glaciers vary between 3700 and 5200 m/s, whereas in superglacial moraines they lie under 3000 m/s. The petrographic nature of the subglacial ground moraine with these extremely high velocities is still unexplained.

The true *longitudinal velocities in the bedrock* range from 5400 to 6000 m/s, with a mean value of 5700 m/s.

The seismologically determined ice thickness differs between 50 and 150 m on the snowfields and 130 and 290 m on the tongues of the glaciers. The total ice and snow mass of the Kesselwandferner amounts to 0.3, that of the Hintereisferner to 0.8, that of the Brandnerferner to 0.027 and that of the Gurglerferner to 0.3 km³. On the Gurglerferner furthermore it was possible to compute the ground moraine volume of the surveyed area to 0.40 km³ and to estimate their total mass to 0.45 to 0.50 km³.

Contour-line maps could be drawn for the surface of the ground moraine and on the Gurglerferner in addition for the bedrock of the valley. They show that the bedrock has several basins and stairs, the morphology of which is marked on the surface of the ground moraine as well as on that of the glacier. At the top of the stairs there is no or little ground moraine. In the front of and behind the steps ground moraine is deposited in great thickness. The superglacial medial moraine is developed as a wall in the subglacial ground moraine, but only above the high stair.

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EFFECT OF HYDROSTATIC PRESSURE ON VELOCITY OF SHEAR DEFORMATIOH OF SINGLE CRYSTALS OF ICE

(Dr.) GEORGE P. RIGSBY

Summary

Apparatus was built for deforming ice crystals under hydrostatic pressures up to 350 atmospheres. Single crystals were placed in the mounts in such a way that the deformation occurred by gliding on the basal glide plane. It was found that the shear strain rate increased as the pressure was increased at constant temperature, but that the rate is practically independent of hydrostatic pressure when the difference between the ice temperature and the melting point is kept constant.

TOME IV

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